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REFRACTION OF OCEAN WAVES: A PROCESS LINKING UNDERWATER TOPOGRAPHY TO BEACH EROSION

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Scripps Institution of Oceanography[1]

ABSTRACT

Waves out at sea, though usually forming a complex pattern, have essentially the same characteristics over large distances. Upon entering shallow water, these waves are transformed under the influence of bottom features, and such transformations may be so marked that breaker heights may vary greatly over short distances along the shore. The effect of bottom features upon waves can largely be interpreted in terms of a simple physical process—wave refraction. In turn, wave refraction may be responsible for alteration of the bottom features by accumulation or removal of sediments and, in this manner, be an important factor in beach erosion.

In this report the role of wave refraction is first reviewed in the light of other processes affecting the transformation of waves in shallow water. The mechanism of refraction is illustrated by means of a few idealized examples, such as the refraction pattern along a straight uniformly sloping beach, over a submarine canyon and ridge, and around a headland. Next it is shown that extreme variations in breaker height along the beach north of La Jolla, California, can be computed for typical swell conditions, taking the complex local bottom topography and the orientation of the coastline into consideration. These changes are computed from refraction diagrams for typical swell conditions, and they compare favorably with observed changes in wave height, thus indicating that wave refraction is the primary mechanism controlling changes in wave height along a beach, and that friction, diffraction, and other processes can be of secondary importance only. Finally, it is noted that the transportation of sediments is dependent upon longshore currents, rip currents, and horizontal diffusion and that all these factors are greatly influenced by the existing refraction pattern.

INTRODUCTION

Waves may have traveled many days and for thousands of miles before they approach shore and become transformed under the influence of the ocean bottom. By that time the long, low, regular undulations, called "swell," may be all but hidden beneath a local chop. As soon as the swell enters shallow water, it "feels" bottom, and a striking rejuvenation takes place. This change occurs in water of depth less than one-half the wave length, where wave length is defined as the horizontal distance between adjacent crests. The wave velocity and length decrease while the height usually increases, so that the wave gradually steepens. As the waves approach the surf zone, this process of steepening accelerates suddenly—the crests rise sharply from the water surface, become peaked, and break.

The changes just described are most noticeable for a swell from a distant storm. Waves formed by a near-by storm, or by winds blowing directly toward the coast, have already considerable steepness out at sea; the transforma-

[1] Contribution from the Scripps Institution of Oceanography, New Series, No. 307. This work represents results of research carried out for the Hydrographic Office, the Office of Naval Research, and the Bureau of Ships of the Navy Department under contract with the University of California.

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tion in shallow water, though similar, is therefore much less pronounced.

Only the period, defined as the time interval between the appearance of two successive crests at a fixed point, remains constant as waves travel into shallow water. According to theory, well substantiated by observation, wave velocity and length depend almost entirely upon wave period and depth (see the Appendix). For a regular wave system the velocity and length will therefore be constant along any given depth contour.

The discussion so far has dealt with changes in the character of the waves which are the direct result of changes in velocity and length. Except under very special conditions, the wave height will, however, not be uniform along any given depth contour, because the effect of refraction over an irregular bottom enters. This effect causes waves to bend, and the bending of the wave crests in turn affects the wave height. With irregular bottom topography, waves are bent differently at different sections of a beach, and the wave height will vary along the beach. The present study is concerned principally with a discussion of the refraction effects.

The refraction of ocean waves and other problems dealing with the transformation of waves in shallow water have been studied intensively during the war in connection with the development of a method for forecasting surf conditions for amphibious operations.[2] Practically all changes in the character of the waves in shallow water can be accounted for theoretically by assuming conservation of energy and irrotational wave motion, and it has been shown that the conclusions based on theory are in good agreement with observations.[3]

WAVE REFRACTION

THE PHYSICAL NATURE OF WAVE REFRACTION

Wave crests approaching a coastline at an angle tend to swing parallel to shore (Fig. 1). This happens because the wave velocity decreases as the depth decreases, so that the portion of the crest nearer the shore moves slowly while the portion of the crest in deeper water races ahead. To illustrate this, consider two surfboarders on the same crest, traveling along the lines A and B (Fig. 1). Surfboarder B always travels faster than A, and the two tend to become more nearly aligned with the shoreline as they travel into shallow water.

The process is similar to the one causing the bending of light rays in optical systems and has therefore been termed "refraction." In the case of straight parallel contours, the analogy can be carried further, because the changes in direction of light waves and ocean waves are governed by Snell's law:[4]

$$\frac{\sin \alpha}{\sin \alpha_0} = \frac{C}{C_0}, \qquad (1)$$

where α is the angle between the wave crest and the contours at any depth, C the wave velocity at the same depth, and the parameters with subscript o refer only to deep water where the direction and velocity are constant.

REFRACTION DIAGRAMS

In the case of a more complicated bottom topography for which the depth contours are not straight and parallel,

[2] "Breakers and Surf, Principles in Forecasting," *Hydrographic Office, U.S. Navy, No. 234* (November, 1944).

[3] H. U. Sverdrup and W. H. Munk, "Theoretical and Empirical Relations in Forecasting Breakers and Surf," *Trans. Amer. Geoph. Union*, Vol. XXVII, No. 6 (December, 1946), pp. 828–36.

[4] Pp. 6–7 of ftn. 2.

changes in wave direction and height can be found graphically by constructing a "refraction diagram," a term first used by O'Brien and Mason.[5]

A refraction diagram is one on which the position of wave crests at the sea surface is shown by a series of lines. These are evenly spaced in deep water—that is, in water of depth greater than half the wave length. Beyond this depth the effect of bottom contours on any of the wave characteristics is negligible. At shallower depths the relation between wave velocity and water depth is known (see Appendix), and a refraction diagram can be constructed by advancing various points on a crest through distances determined from any chosen time interval and the average depth. The direction of advance is drawn normal to the crest. In Figure 1, for example, the chosen time interval is one wave period, and the distance A_0—A'. The distance B_0—B' represents the advance of point B during the same time interval, but for the average depth between B_0 and B'. The lines A_0—A' and B_0—B' are normal to the wave crests. By locating a set of points, A', B', C', ... , the new crest can be found with sufficient accuracy by drawing a smooth line through these points. Advancing one wave length at a time, the crests may be carried into any depth of water desired, and the completed diagram shows the continuous change in the direction of a wave advancing from deep into shallow water.

Description of a practical procedure for constructing refraction diagrams, together with the necessary graphs and tables, can be found in the forecasting manual.[6] The completed diagram can be interpreted in one of two ways: (1) As a series of lines representing the positions of a *single* wave crest at various times as the crest advances toward shore. Crest interval is then defined as the interval between these times. In Figure 1 the crest interval equals one wave period. (2) As a series of lines representing the position of certain wave crests at a *single instant;* a crest interval of *one* wave length means that every crest is represented (Fig. 1), because the advance of any point on a crest during one wave period equals one wave length by definition. A crest interval of *two* wave lengths means that only every second wave crest is shown; a crest interval of *one-half* wave length, that the position of every crest and trough is shown.

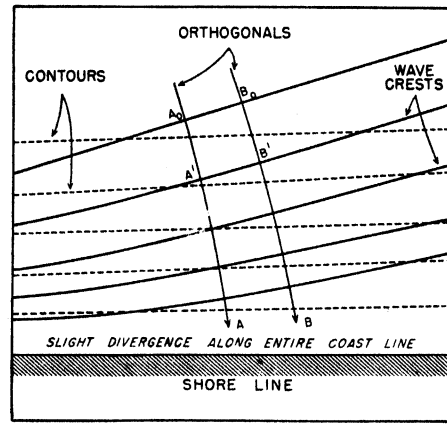

FIG. 1.—Refraction of waves along beach with straight parallel depth contours. The wave crest along the orthogonal B is always in deeper water than along A and therefore moves with greater velocity. As a result, the waves tend to turn parallel to shore.

EFFECT OF REFRACTION ON WAVE HEIGHT

Wave height is defined as the vertical distance between crest and trough. In order to evaluate the effect of refraction on wave height, a set of orthogonals, that

[5] "A Summary of the Theory of Oscillatory Waves," *Technical Report No. 2, Beach Erosion Board* (1940).

[6] Pp. 31–34 of ftn. 2.

is, a family of lines which are everywhere perpendicular to the wave crests, must be constructed. The lines A_0–A and B_0–B in Figure 1 are orthogonals; they may be visualized as the wakes behind two surfboards which are continuously oriented normal to the crest, that is, in the direction of wave motion. Assuming that the wave *energy* is transmitted in the direction of wave motion, the total flow of energy between two orthogonals must remain constant. Thus if orthogonals *converge*, the crests are *compressed*, and the energy per unit crest length is relatively *large;* if orthogonals *diverge*, the crests are *stretched*, and the energy per unit crest length is relatively *small*.

For practical purposes one must deal with wave heights rather than with wave energies, but these are related. In deep water the height of oscillatory waves varies as the *square root* of energy, and this relation holds with good approximation in shallow water until the waves "peak" just prior to breaking. The breaking waves, with their sharp crests isolated by long flat troughs, have the appearance of so-called "solitary" waves, for which the height is proportional to the *cube root* of the energy. Let H designate the wave height and s the distance between adjacent orthogonals on the refraction diagram. Parameters with subscript o again apply in deep water, those with subscripts b at the breaker point, and parameters without subscript at any depth intermediary between deep water and the breaking-point. Then

$$\frac{H}{H_0} = \gamma K, \quad \frac{H_b}{H_0} = \gamma_b K_b, \quad (2a, b)$$

where γ and γ_b have constant values along a fixed depth contour, and where

$$K = \sqrt[2]{s_0/s}, \quad K_b = \sqrt[3]{s_0/s_b}, \quad (3a, b)$$

may vary along a depth contour and will be referred to as the "refraction factors." The complete derivation of these equations is given in the Appendix.

Any *variation* in wave height along a fixed depth contour depends only upon the refraction factor. The variation can be computed from the change in distance between orthogonals according to equations (3), depending upon whether one deals with conditions in shallow water or along the breaker line. Again it should be stressed that the absolute wave height depends also upon several other factors which are implicitly contained in the parameters γ and γ_b of equations (2) (see Appendix).

Equations (2) are derived from the postulates that the energy flows along orthogonals and that energy is conserved. This in turn implies two assumptions: (1) the effect of *diffraction* (which would bring about flow of energy across orthogonals from region of high waves to regions of low waves) can be neglected and (2) the effect of bottom friction is negligible. The validity of these assumptions is borne out by the good agreement between computed changes in breaker height according to equations (2), and observations along the beach north of La Jolla, California (Fig. 16). There has been a tendency in the literature to emphasize the effect of bottom friction on wave motion and other wave characteristics. Not only the evidence presented in this paper, but also studies dealing with the generation of waves by wind,[7] and with absolute changes of wave height in shallow water and the dynamics of breaking waves,[8] indicate that wave motion in general is not appreciably affected by frictional processes.

[7] H. U. Sverdrup and W. H. Munk, "Empirical and Theoretical Relations between Wind, Sea and Swell," *Trans. Amer. Geoph. Union*, Vol. XXVII, No. 6 (December, 1946), pp. 823–27.

[8] Pp. 830–32 of ftn. 3.

REFRACTION TYPE EXAMPLES

Beach with straight and parallel depth contours.—For this simple case the change in wave direction can be expressed analytically by equation (1). Figure 1 is a schematic drawing of wave refraction along a straight coastline with parallel depth contours. Plate I, *A*, is an aerial photograph showing long waves from the south-southwest coming into the beach north of Oceanside, California. There is a divergence of the orthogonals indicated by the values of s computed from the equation[9]

$$\frac{s}{s_0} = \frac{\cos \alpha}{\cos \alpha_0}. \qquad (4)$$

[9] Pp. 20–21 of ftn. 2.

Combining equations (1), (3), and (4) leads to

$$\left.\begin{aligned} K &= \left[\frac{\cos^2 \alpha_0}{1-\left(\frac{C}{C_0}\sin \alpha_0\right)^2}\right]^{1/4}, \\ K_b &= \left[\frac{\cos^2 \alpha_0}{1-\left(\frac{C_b}{C_0}\sin \alpha_0\right)^2}\right]^{1/6}, \end{aligned}\right\} \qquad (5a, b)$$

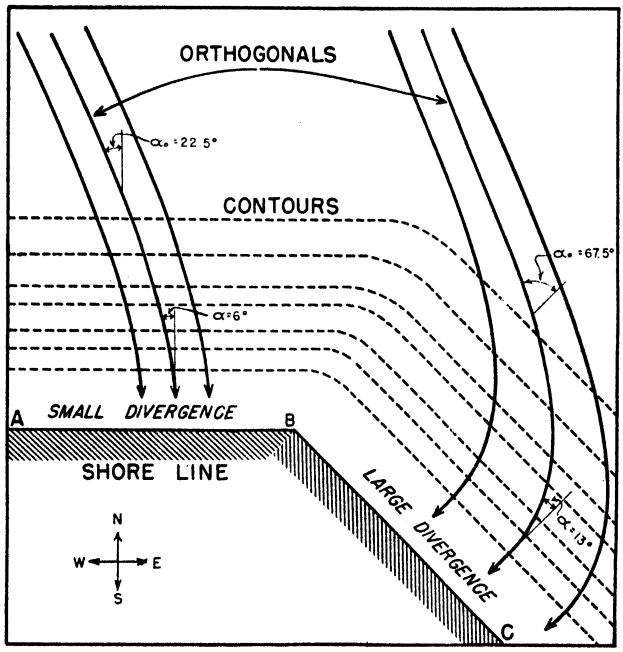

FIG. 2.—Refraction of waves along two sections of straight beaches with different exposures. The divergence along section *A–B* is less than along *B–C*, and the wave height along *A–B* exceeds therefore the wave height along *B–C*.

where C/C_0 is found from equation (A7) in the Appendix.

Waves coming from an angle into a beach with straight and parallel depth contours are reduced in height according to equations (5), but the reduction is uniform along the entire beach, and no *variation* in wave height *along* the beach will result. As an example of variation in wave height, consider a coastline which

forms a sharp bend at point *B* (Fig. 2). On both sides of point *B* the coastline and the depth contours are straight and parallel. Assume that waves of 14-second period come from the north-northwest. The angle α is drawn between the orthogonals and the bottom gradient, which is consistent with our earlier definition. Table 1 gives the computed direction and relative height at the 10-foot contour, which is assumed to lie outside the breaker zone. Along the beach section *A–B*, which is relatively exposed to the incoming swell, waves are about 50 per cent higher than along section *B–C*. This can also be seen from the fact that the divergence of orthogonals along *A–B* is less than the divergence along *B–C* (Fig. 2).

Submarine canyons.—Portions of wave crests over the center of a canyon are in deeper water and move ahead faster than the portions on either side (Fig. 3). Consequently, the waves fan out, resulting in divergence (low waves) over the head of the canyon and convergence (high waves) on either side. Fishermen take advantage of this feature when they anchor near the head of a submarine canyon. Plate II is an aerial photograph of waves over the submarine canyon at Redondo, California. The approximate position of the canyon is indicated. The wave crests bend sharply over the canyon walls. Variations in wave height can be recognized by the prominence of the wave crests on either side of the canyon

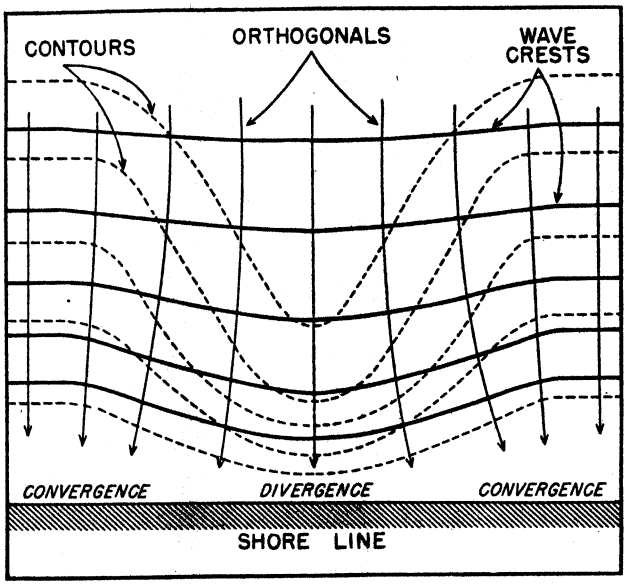

FIG. 3.—Refraction of waves by a submarine canyon. Waves move faster over the canyon than on either side of the canyon, resulting in divergence (low waves) over the mouth of the canyon and convergence (high waves) on either side.

TABLE 1

Section of Coast Line (Fig. 2)	A–B	B–C
Wave angle in deep water (α_0)	22.5°	67.5°
Wave angle at 10-foot depth (eq. [1])*	6.0°	13.0°
Refraction factor (eq. [5a])	0.97	0.63

* The value of C/C_0 is obtained from "Breakers and Surf," *op. cit.*, Pl. I, and equals 0.25.

compared to the crests directly over the canyon and by the variation in the width of the surf zone.

Submarine ridges.—Underwater ridges near shore have an effect opposite to that of canyons. Waves passing over the ridge are in shallower water and are therefore retarded, and on either side waves move ahead, creating a convergence over the ridge (Fig. 4). This is borne out by frequent observations of unusually violent breakers over the shoal portion of underwater ridges. An aerial photograph take over the northern shore of Oahu, T.H. (Pl. I, *B*), illustrates the lagging of the wave crests over a submarine ridge and the resulting convergence. Figure 5 illustrates the same features over a sand bank opposite the mouth of Mission Bay, San Diego, California. In the case of Mission Bay the refraction pattern is somewhat affected by the tidal flow through the outlet.

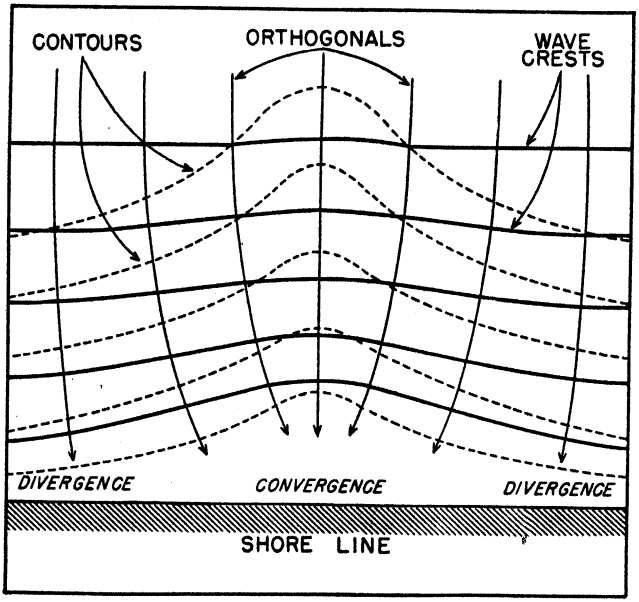

FIG. 4.—Refraction of waves by a submarine ridge. Directly over the ridge waves lag behind, and on either side waves move ahead, creating a convergence (high waves) over the ridge.

Abrupt changes in depth along coral islands lead to particularly well-defined refraction patterns. Plate III shows two channels in the coral reef around Guam; the waves are seen to bend sharply ahead over the channels, where they are lower than on either side of the channels. The photograph also shows the formation of multiple crests just inside the reef, a phenomenon often observed in regions of abrupt depth changes, and one which has never been properly explained.

Headlands.—Since waves are retarded by shallow water, they will bend around headlands, points, and jetties. Plate IV, *A*, shows an irregular train of waves turning around Point Loma into San Diego Bay.

Under special conditions, such as portrayed in the aerial photograph of the hook of Cape Cod, Massachusetts (Pl. IV, *B*), waves may bend 180°, or even more, before they break. Such extreme

forms a sharp bend at point *B* (Fig. 2). On both sides of point *B* the coastline and the depth contours are straight and parallel. Assume that waves of 14-second period come from the north-northwest. The angle α is drawn between the orthogonals and the bottom gradient, which is consistent with our earlier definition. Table 1 gives the computed direction and relative height at the 10-foot contour, which is assumed to lie outside the breaker zone. Along the beach section *A–B*, which is relatively exposed to the incoming swell, waves are about 50 per cent higher than along section *B–C*. This can also be seen from the fact that the divergence of orthogonals along *A–B* is less than the divergence along *B–C* (Fig. 2).

Submarine canyons.—Portions of wave crests over the center of a canyon are in deeper water and move ahead faster than the portions on either side (Fig. 3). Consequently, the waves fan out, resulting in divergence (low waves) over the head of the canyon and convergence (high waves) on either side. Fishermen take advantage of this feature when they anchor near the head of a submarine canyon. Plate II is an aerial photograph of waves over the submarine canyon at Redondo, California. The approximate position of the canyon is indicated. The wave crests bend sharply over the canyon walls. Variations in wave height can be recognized by the prominence of the wave crests on either side of the canyon

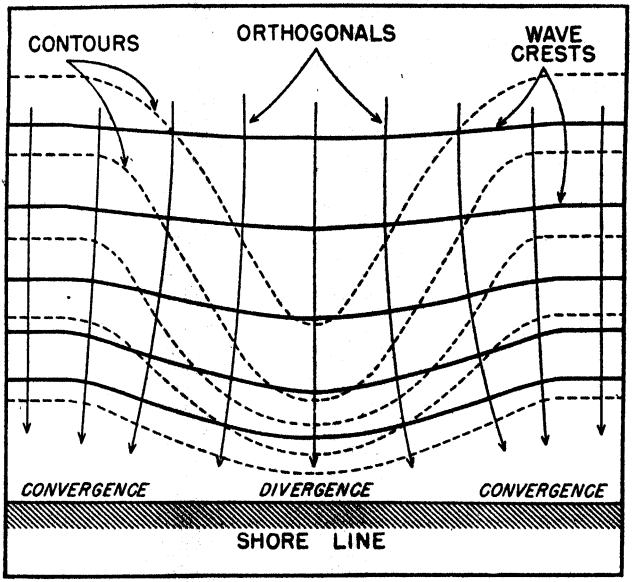

FIG. 3.—Refraction of waves by a submarine canyon. Waves move faster over the canyon than on either side of the canyon, resulting in divergence (low waves) over the mouth of the canyon and convergence (high waves) on either side.

TABLE 1

Section of Coast Line (Fig. 2)	A–B	B–C
Wave angle in deep water (α_0)	22.5°	67.5°
Wave angle at 10-foot depth (eq. [1])*	6.0°	13.0°
Refraction factor (eq. [5a])	0.97	0.63

* The value of C_1/C_0 is obtained from "Breakers and Surf," *op. cit.*, Pl. I, and equals 0.25.

compared to the crests directly over the canyon and by the variation in the width of the surf zone.

Submarine ridges.—Underwater ridges near shore have an effect opposite to that of canyons. Waves passing over the ridge are in shallower water and are therefore retarded, and on either side waves move ahead, creating a convergence over the ridge (Fig. 4). This is borne out by frequent observations of unusually violent breakers over the shoal portion of underwater ridges. An aerial photograph take over the northern shore of Oahu, T.H. (Pl. I, *B*), illustrates the lagging of the wave crests over a submarine ridge and the resulting convergence. Figure 5 illustrates the same features over a sand bank opposite the mouth of Mission Bay, San Diego, California. In the case of Mission Bay the refraction pattern is somewhat affected by the tidal flow through the outlet.

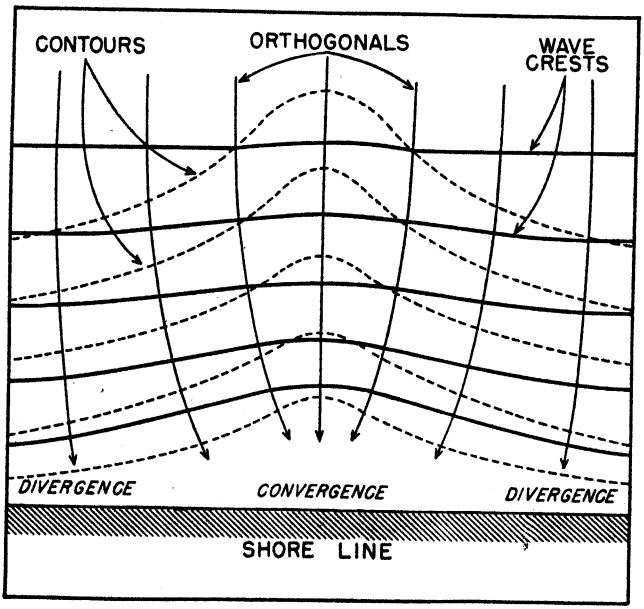

FIG. 4.—Refraction of waves by a submarine ridge. Directly over the ridge waves lag behind, and on either side waves move ahead, creating a convergence (high waves) over the ridge.

Abrupt changes in depth along coral islands lead to particularly well-defined refraction patterns. Plate III shows two channels in the coral reef around Guam; the waves are seen to bend sharply ahead over the channels, where they are lower than on either side of the channels. The photograph also shows the formation of multiple crests just inside the reef, a phenomenon often observed in regions of abrupt depth changes, and one which has never been properly explained.

Headlands.—Since waves are retarded by shallow water, they will bend around headlands, points, and jetties. Plate IV, *A*, shows an irregular train of waves turning around Point Loma into San Diego Bay.

Under special conditions, such as portrayed in the aerial photograph of the hook of Cape Cod, Massachusetts (Pl. IV, *B*), waves may bend 180°, or even more, before they break. Such extreme

refraction is associated with a very large divergence, which explains the shelter usually provided by protruding land masses.

Bays and islands.—Waves will be retarded along the sides of a bay but move relatively rapidly elsewhere. As a result, the crests will bend *into* the bay (Pl. V). The divergence will be largest at the sides of the bay, least opposite the mouth, especially if waves head directly into the bay. In the case of small islands in the path of an advancing swell the resulting refraction pattern is more complicated (Fig. 6). Offshore in the lee of the island, wave trains which have bent around both sides of the island meet and pass through each other, and a confused cross-sea results.[10] Where the bottom shoals gradually near shore, then drops off steeply, waves may be bent so effectively around an island that the lee shore offers no protection whatsoever.[11] Some coral islands have these characteristics.

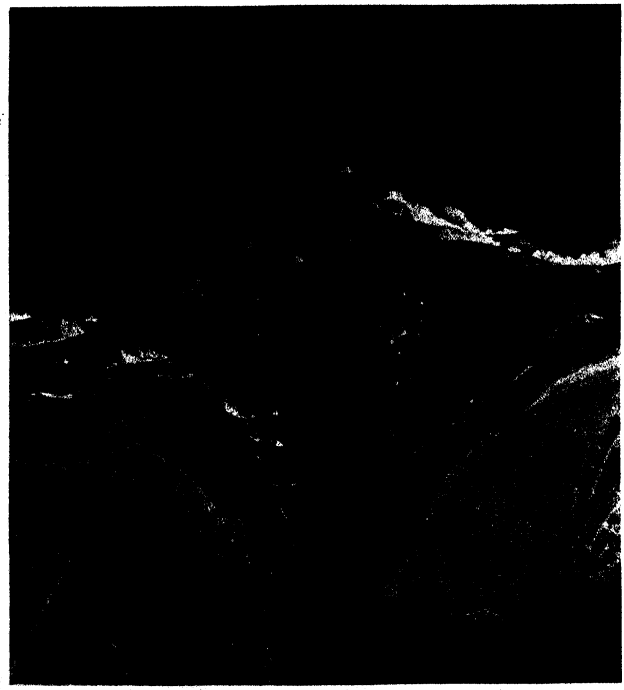

FIG. 5.—Refraction of waves over sand bar outside of Mission Bay, San Diego, California (cf. Fig. 4) The outflowing tide may have further strengthened certain features in the existing refraction pattern. (Official photograph, United States Navy.)

WAVE REFRACTION ALONG THE BEACH NORTH OF LA JOLLA, CALIFORNIA

ORIGIN OF WAVES REACHING LA JOLLA

In order to apply the principles of refraction to an actual situation, it is first necessary to determine the periods and directions of waves that will be encountered. For that purpose the character of the waves at La Jolla resulting from typical meteorological situations will be tabulated below. The paths and

[10] Pp. 12–18 of ftn. 2.

[11] Robert S. Arthur, "Refraction of Water Waves by Islands and Shoals with Circular Bottom-Contours," *Trans. Amer. Geoph. Union*, Vol. XXVII (April, 1946), pp. 168–77.

nature of the generating areas are shown schematically in Figure 7; the corresponding character of the waves at La Jolla, in Figure 7, *B*.

Situation A (*winter*).—Waves generated by cyclones originating at the east coast of Asia and moving along the usual cyclone path into the Gulf of Alaska:

Waves generated in the southern segments of these cyclones reach La Jolla in 2–5 days as a long-period (11–15 seconds) swell, with heights varying between 3–7 feet, according to the intensity of the cyclone. The wave energy reaching La Jolla is reduced by the offlying islands, especially San Clemente and Santa Catalina (Fig. 7, *A*). The cyclones often stagnate in the Gulf of Alaska, and the coasts of British Columbia, Washington, Oregon, and northern California will be hit by a heavy swell, whereas southern California will be protected by Point Conception, which cuts off La Jolla from waves from directions much north of WNW.

Situation B (*winter*).—Waves generated behind cold fronts that come into the coastal regions:

These waves have short periods (7–10 seconds), large heights (4–12 feet), and usually come from a W to NW direction.

Situation C (*summer*).—Waves generated by winds in the northeast periphery of the Pacific High:

Usually the period of the waves will vary from 6 to 9 seconds, the height from 2 to 5 feet. The common direction is from WNW to NW.

Situation D (*summer*).—Waves generated in the Southern Hemisphere:

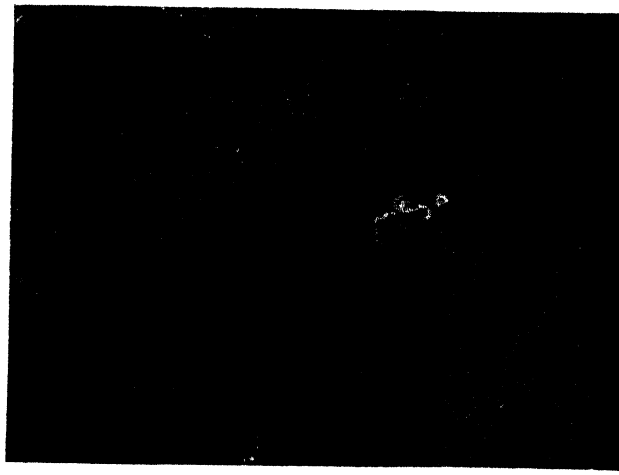

FIG. 6.—Refraction of waves behind a small island. The portion of the wave passing just to the right of the island is refracted to the left; the portion passing just to the left of the island is refracted to the right. The resulting crisscross pattern gives rise to a confused sea in the lee of the island, and the island may offer little or no shelter.

The so-called "southern swell" comes from S to SSW, is of very long period (13–20 seconds), and is usually 3–5 feet high. This is the only direction from which waves can reach La Jolla without being broken up by offlying islands.[12] There is strong evidence that this swell has been generated by the winter storms of the Southern Hemisphere, but, owing to an almost complete lack of weather observations, this has not been verified.

Situation E (*spring*).—Wind waves generated in local low-pressure regions:

Local low-pressure regions, which are usually not forecast, give rise to W or NW winds with

[12] Plate I, *A*, is an aerial photograph of a southern swell coming into Oceanside, Calif.

velocities up to 30 knots. These winds raise waves of 5–7 second-periods, 10–15 feet-heights. Such situations are likely to occur a number of times each spring, occasionally during winter and summer, but rarely during fall.

The character of the waves caused by these five meteorologic situations is summarized in Figure 7, *B*. Each rectangle shows typical height-period combinations corresponding to situations *A–E*.

LOCAL TOPOGRAPHY

The coast between Point La Jolla and the Scripps Institution pier is ideally suited for a study of changes in wave height. It is known locally that waves are usually low near the Beach Club, then increase sharply about 3,000 feet north of the Beach Club, and fall off a little in the vicinity of the Scripps Institution pier (Pl. VI). Breaker heights at different sections of this 1-mile stretch of beach generally differ by a factor of 3.

Such unusually large variations in breaker height can be associated with the complex submarine topography and coastline in this region. Two submarine canyons, La Jolla Canyon and its steep-walled tributary, Scripps Canyon, reach within 1,000 feet of shore (Pl. VI). These canyons have been sounded more extensively than any other submarine canyons in the world.

The walls of La Jolla Canyon are relatively gentle except locally where slopes of 100 per cent are found. Scripps Canyon on the other hand has extremely steep walls. Numerous places have been located where the slopes are so steep that the sounding lead slides down the slope after hitting bottom. The steepest measured slope has a drop of 230 feet in a horizontal distance of 30 feet. The canyon has an inner gorge which is so narrow in places that it was difficult to locate. . . . .[13]

The coast line, which is oriented NNE–SSW between Scripps pier and the Beach Club, makes an almost right-angle turn 1,000 feet south of the Beach Club and projects about 3,000 feet in a WNW direction. Point La Jolla is located at the end of this break in the coast line (Pl. VI). Beyond Point La Jolla the coast continues its NNE–SSW trend. In the vicinity of the Beach Club some protection from the usual WNW and W swell is afforded by Point La Jolla, and even more by the shoal area off this point.

TABLE 2

	Figure No.				
	8	9	10	11	12
Wave direction	WNW	WNW	W	WNW	SSW
Wave period	14	12	12	8	16
Crest interval/period	1	1	1	2	1
Meteorological situation	A	A	A	B, C, or E	D

REFRACTION DIAGRAMS

The refraction diagrams (Figs. 8–12) were drawn as described in the manual on breaker and surf forecasting. Table 2 gives the assumed wave direction (direction from which the wave comes in deep water), wave period, crest interval, and the corresponding meteorological situation for each of the five figures.

[13] F. P. Shepard and K. O. Emery, "Submarine Topography off the California Coast," *Geol. Soc. Amer., Special Papers, No. 31* (May 28, 1941).

The thin dashed lines on the figures are bottom contours in fathoms; the heavy solid lines are wave crests; and the light solid lines with arrowheads are orthogonals drawn at arbitrary intervals. For waves with periods more than 10 seconds every single crest has been drawn; for waves with periods less than 10 seconds only every second crest has been drawn.

The area offshore has been divided into divergent, normal, and convergent zones, according to whether the refraction factor K (eq. [3*a*]) is less than 0.75, 0.75–1.25 inclusive, or larger than 1.25. Divergent zones are dotted, convergent zones crosshatched, and normal zones left blank.

The value of K_b (eq. [3*b*]) at various stations is proportional to the length of the rectangles at those stations and can be read off from the scale given in the legend of each refraction diagram. In order to emphasize changes in K_b, the columns have been shaded in the same manner used for the offshore area. Columns *ending* in a dotted section designate divergence (K_b less than 0.75); those ending in a crosshatched section designate convergence (K_b larger than 1.25). It should be kept in mind that values of K are proportional to the square root and values of K_b to the cube root of the relative spacing of the orthogonals. The stations where observations were made are numbered *1–12*. Other stations, designated by the letters *A–K*, have been selected because of their representative location.

Waves from WNW, period 14 seconds (Fig. 8).—A small region of relatively high waves, or convergence, is found just north of Station A. An area of intense convergence, which extends approximately northward from the Cove (Station B), is caused by the steep south wall of La Jolla Canyon. This wall is particularly effective because it lies parallel to the direction of wave travel for half a mile. The prolonged effect of the steep bottom slope caused the waves to turn sharply to the right and converge into the Cove. This probably explains the occurrence of unusually high waves at the Cove on days when waves are low along other sectors of the beach. It is well known in La Jolla that these high waves, on days when they occur, come into the Cove at intervals of 15 minutes to several hours. Perhaps the effect of La Jolla Canyon is particularly marked for certain critical combinations of wave period and direction. Although both period and direction are nearly constant for any wave train, they do show continual small variations, and certain combinations of these periods and directions may be the cause of intermittent high waves at the Cove.

The submarine canyon offers protection along the beach from Station C to Station 5 by creating a divergence. This divergence is particularly marked in the vicinity of the Beach Club, where K_b equals 0.3. A slight convergence in the vicinity of Station 6 is the result of waves turned left by the north bank of La Jolla Canyon and is analogous to the convergence near the Cove, caused by the south bank of the canyon. The divergence between Stations 8 and I is caused by the Scripps Canyon. It is less intense than the divergence due to the La Jolla Canyon because for the direction assumed waves do not travel in a line parallel to the walls of the Scripps Canyon. To the north of the Scripps Canyon, between Stations I and J, a small but sharp convergence is found.

The variation of K_b along the beach is also shown in Figure 15. The lower portion of the figure gives the location of the stations and the two submarine canyons.

The upper figure contains a curve for K_b for each of the refraction diagrams. Low wave heights near the head of the two canyons and high waves to either side are clearly shown.

Waves from WNW, period 12 seconds (Fig. 9).—The general features are similar to those caused by the 14-second WNW waves, but the divergence is less intense and covers a smaller area. Wave heights between the two submarine canyons (Stations 4–9) are relatively higher than for the 14-second wave. This difference is chiefly due to the fact that the 12-second wave is shorter than the 14-second wave and that canyons and other bottom features do not affect the 12-second waves to the same degree that they affect the 14-second waves.

Waves from W, period 12 seconds (Fig. 10).—The main features are similar to the ones already discussed, except that regions of divergence and convergence have been shifted due to the difference of wave direction. Thus the convergence on the south edge of La Jolla Canyon has shifted from the Cove to Station C, and the convergence between the canyons from Station 6 to Station 8. The convergence near Station A is more pronounced than for a west-northwesterly swell. Scripps Canyon gives rise to more extreme divergence between Stations 11 and I because the direction of wave travel in deep water is more nearly parallel to the walls of the canyon.

Waves from WNW, period 8 seconds (Fig. 11).—The length of these waves in deep water is only one-third that of the 14-second swell, and the effect of refraction is reduced even more than it was for the 12-second waves. In particular, regions of divergence have become much smaller, and only about 3,000 feet of beach are protected by the La Jolla Canyon. The region of divergence opposite the Scripps Canyon has disappeared altogether. The convergence on the south side of La Jolla Canyon has moved to Station D, thus eliminating the cause of relatively high waves at or near the Cove. Nevertheless, it can be seen from Figure 15 that the refraction pattern typical of a submarine canyon has not disappeared: waves are relatively low opposite the mouth of the canyons and high on either side.

Waves from SSW, period 16 seconds (Fig. 12).—A few selected orthogonals between deep water and shore have been entered in an inset on a smaller scale in order to show that for this particular case the effect of refraction outside the limits of the main figure cannot be neglected. By far the larger portion of Figure 12 shows values of K less than 0.75. This divergence is due to the shoal area off Point La Jolla. At only two localities in the vicinity of La Jolla can a convergence of a southern swell be found, opposite the Casa de Mañana Hotel (Station A) and to the north between Stations J and K. A divergence followed by a convergence is caused by the ridge between the submarine canyons. It is believed that this explains the occasional occurrence of relatively high, long-period waves between Stations 7 and 9 during the summer. The refraction diagram explains also why, during the summer, waves are relatively low at the Scripps Institution while they are high at localities with southern exposure, such as near Oceanside and the Coronado Strand.

OTHER METHODS OF PRESENTING REFRACTION DIAGRAMS

If an intensive study is made of a given locality, it may be of advantage to construct a single graph from which the refraction factor K_b and the breaker

angle α_b can be read off directly. In general, K_b and α_b depend upon three variables—the wave direction in deep water, the wave period, and the depth of breaking—but K_b and α_b vary slowly with the depth of breaking, and an average value for the breaking depth may be introduced.

A convenient form of presentation is a polar-coordinate graph with its center placed on an outline map at the position to which the graph applies (Fig. 13). The radial lines give the direction from which the waves come in deep water; the circles give the wave period. The solid curves are lines of equal K_b, the dashed curves lines of equal α_b. The "L = LEFT" and "R = RIGHT" refer to an observer on the beach looking straight out to sea.

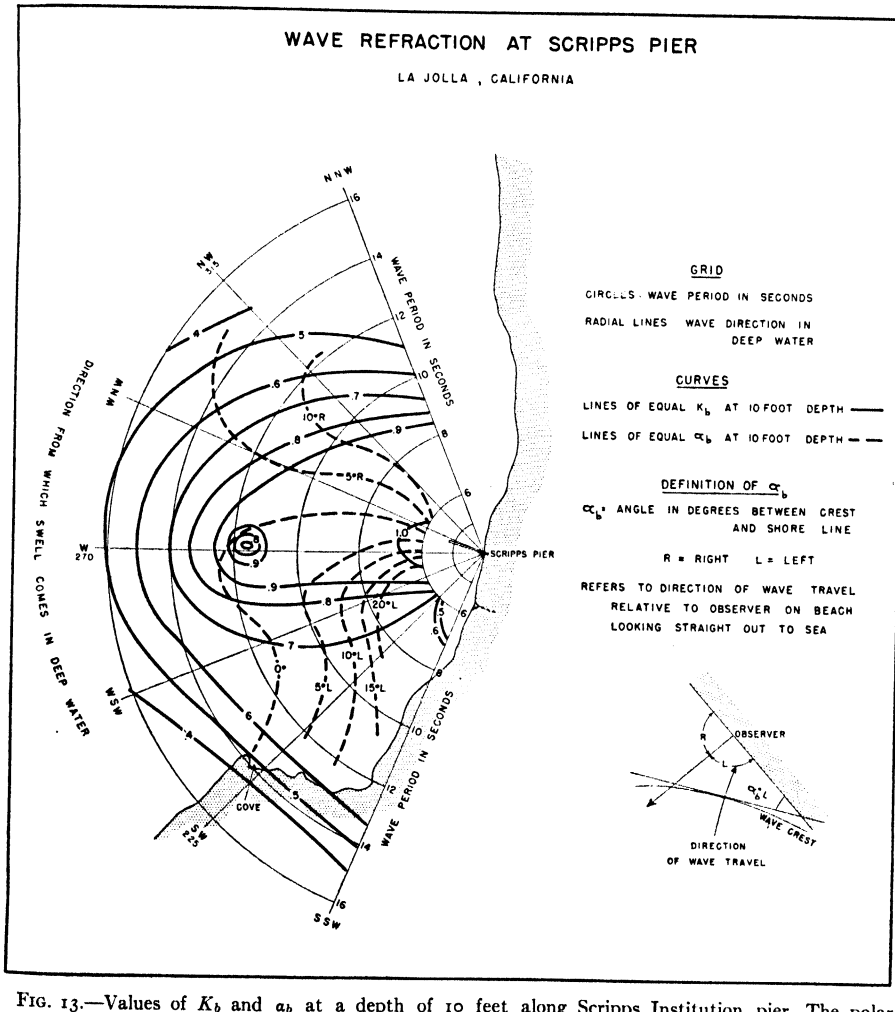

FIG. 13.—Values of K_b and α_b at a depth of 10 feet along Scripps Institution pier. The polar-coordinate presentation brings out the relative protection of the designated location from the waves coming from SSW to WSW, and the relative exposure to waves from the west.

Figure 13 is drawn for an assumed breaking depth of 10 feet along the Scripps Institution pier. The figure applies with good approximations to breakers anywhere between shore and a depth of 20 feet. The curves were drawn by interpolating between values read off Figures 8–12 and a number of other refraction diagrams drawn specially for the region in the immediate vicinity of the pier. The curve $\alpha_b = 0$ refers to breakers coming straight into shore, that is, from a WNW direction. The other dashed curves are drawn relative to this direction.

Figure 14 applies to the case of a beach with straight and parallel depth contours but is otherwise analogous to Figure 13. For this idealized case the curves for K_b and α_b are given analytically by equations (1) and (5*b*). From a comparison between Figures 13 and 14, and assuming $T = 14$ seconds, the special effects of the bottom topography and of the con-

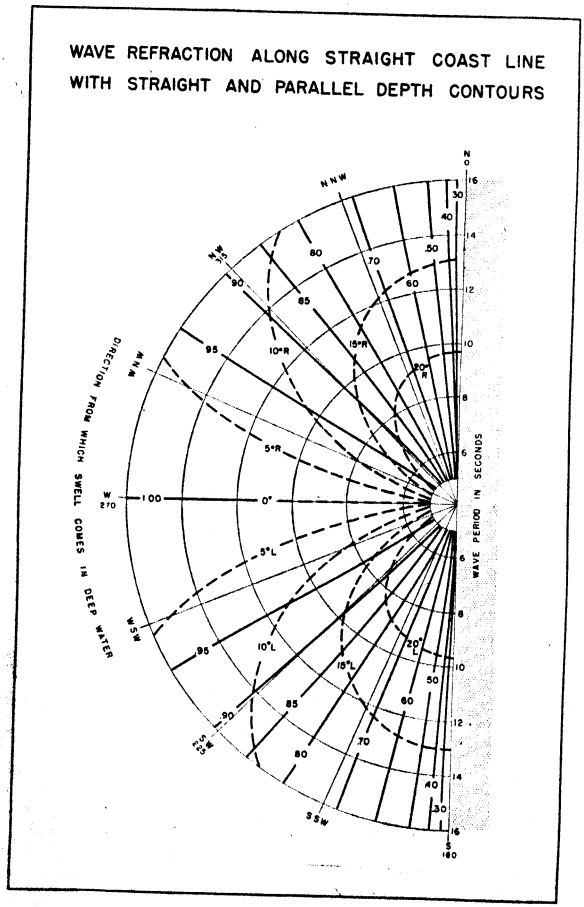

FIG. 14.—Values of K and α_b at a depth of 10 feet for a beach with straight and parallel depth contours (eqs. [1] and [5*b*]). The special effect of the coastline and the bottom topography at La Jolla can be estimated from a comparison between Figs. 13 and 14.

figuration of the coastline upon the breakers near the Scripps pier can be evaluated (Table 3).

Thus the beach along the pier is particularly well sheltered from southerly waves due to the protection afforded by Point La Jolla and is most exposed to waves directly from the west. This type of presentation gives, therefore, an over-all picture of the protection from, or the exposure to, waves of various directions and periods. The disadvantage of this presentation is that it is limited to a relatively small area. If the emphasis is to be placed upon the changes in wave characteristics along a beach for a given wave situation, the presentation shown in Figure 15 will be more practical.

TABLE 3

	Fig.	Direction from Which Waves Come in Deep Water					
		SSW	SW	WSW	W	WNW	NW
Refraction factor K_b	13	0.51	0.45	0.57	0.70	0.60	0.45
	14	0.72	0.90	0.97	1.00	0.97	0.90
Breaker angle α_b	13	.....	2°L	1°R	2°R	4°R	7°R
	14	13°L	10°L	4°L	0°	4°L	10°L

OBSERVATIONS

Naval officers under instruction at the Scripps Institution, after having been trained for one week in a uniform procedure of observing waves, occupied Stations 1–12 along the beach and made simultaneous observations for 15 minutes. These observations were averaged to obtain observed values. To minimize differences due to personal judgment, observers were shifted regularly from one station to another. From about seventy complete sets of observations, 7 days were selected during which the swell period was well defined and in agreement with forecasts. According to forecasts, all these waves came from WNW. Observations are summarized in Table 4. For each of the twelve stations Table 4 gives the height of the breakers in feet, the height of the breakers as percentage of the average for all stations, and the wave period in seconds. The value for H_b at Station 6 on January 5 appears to be too high, but otherwise observations are fairly consistent.

For each *station* observations made on the 7 days under consideration were averaged and entered in the last column of Table 4. For each *day* observations made at the twelve stations were averaged and entered at the bottom of the table, together with the forecast values. On the 7 days under consideration the average periods vary between 11.5 and 13.8 seconds. The over-all average for observed periods is 12.5 seconds, and for the forecast periods 14 seconds. The forecast direction is WNW. Figures 8 and 9 should therefore apply in all seven cases.

COMPARISON OF THEORY AND OBSERVATIONS

The *computed* refraction factor K_b for waves of 14- and 12-second periods from the WNW was read off at each of the twelve stations. In the case of the 14-second wave, the average value for all stations was 0.62; in the case of the 12-second wave, 0.82. At each station the

individual values of K_b were then expressed as percentages of these average values and plotted in Figure 16 (solid and dashed curves).

The dots in Figure 16 give the average *observed* breaker height and the crosses the average *observed* wave periods, both expressed as percentages of the over-all average for all days and stations. These values were taken from the last column of Table 4. In view of the inaccuracies of the observations and the difficulty in drawing exact refraction diagrams, the agreement between theory and observation is satisfactory. The only noticeable discrepancy occurs near Station 7, where the computed wave height exceeds the observed height by about 10 per cent. This discrepancy may be related to a shifting sand bank about 2,500 feet south of the pier.

According to theory, the wave periods should remain constant. Observations plotted in Figure 16 show that this is nearly the case, the lowest value being 95, and the highest 105, per cent of the average. It happens that these small variations in period are in phase with the variations in wave height.

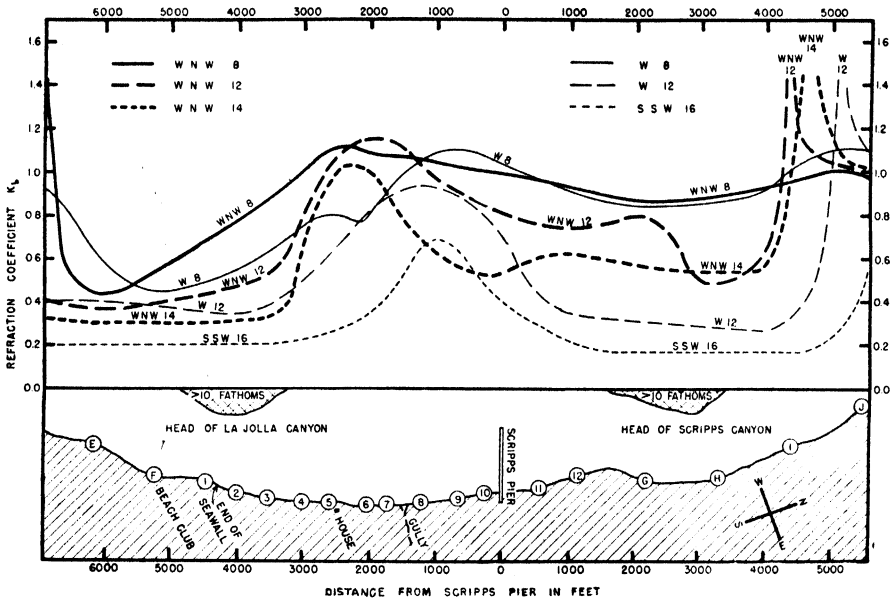

FIG. 15.—Variations of K_b along the coastline according to the refraction diagrams. The lower section of the figure shows the locations along the beach to which values of K_b in the upper section refer. For all but the southern swell, waves are generally lower near the head of the canyon than on either side. Short wave periods give smaller variations in wave height than longer periods. Changes in wave direction shift the location of the zones of convergence and divergence, but they have little effect upon the amplitude of the variation in K_b.

The refraction diagrams can also be compared directly to aerial photographs. Plate VII was selected for an example, because it shows an unusually regular, well-defined swell. The area covered in the photograph extends between Station 5 and Station G.

The swell shown on the photograph was generated in a low-pressure area

TABLE 4

OBSERVATIONS OF BREAKER HEIGHT AND PERIOD

Sta. No.	Dec. 1, 1944			Dec. 8, 1944			Jan. 2, 1945			Jan. 3, 1945			Jan. 5, 1945			Feb. 13, 1945			Feb. 14, 1945			Per Cent Averages	
	H_b		T (Sec.)	H_b		T (Sec.)	H_b		T (Sec.)	H_b		T (Sec.)	H_b		T (Sec.)	H_b		T (Sec.)	H_b		T (Sec.)	H_b	T
	Ft.	C_c		Ft.	C_c		Ft.	C_c		Ft.	C_c		Ft.	C_c		Ft.	C_c		Ft.	C_c			
1	0.7	33	11.5	2.0	38	11	1.0	45	.....	1.0	34	14.0	1.0	22	12.0	3.0	75	12.0	1.5	50	10.0	42	95
2	1.3	62	12.0	2.0	38	11	1.0	45	14.5	1.0	34	11.0	1.5	33	14.0	2.5	63	12.0	2.0	67	9.0	49	94
3	1.5	71	12.0	.....	.....	.....	2.5	114	12.0	1.5	52	13.5	2.0	44	14.0	1.8	45	14.0	3.0	100	12.0	71	102
4	1.5	71	12.0	5.5	104	12	2.5	114	14.0	3.5	121	11.0	4.0	87	12.0	5.0	125	12.0	3.0	100	12.0	103	97
5	2.5	119	13.0	6.0	113	12	3.0	137	14.0	5.0	173	14.0	6.0	130	14.5	6.0	150	13.0	4.0	133	12.0	136	105
6	3.5	167	12.0	4.5	85	12	3.5	159	14.5	4.0	138	11.0	11.0	240	16.0	4.5	113	12.0	4.0	133	11.0	148	103
7	3.5	167	12.0	5.0	94	13	2.5	114	14.0	4.0	138	15.0	5.0	109	15.0	5.0	125	12.0	3.5	117	11.0	123	104
8	3.0	143	12.0	6.0	113	12	2.0	91	13.0	2.5	86	10.0	4.5	98	13.0	4.0	100	12.5	3.5	117	14.0	107	99
9	2.5	119	13.0	6.5	123	13	2.5	114	13.0	3.5	121	11.0	4.5	98	13.0	3.5	88	11.5	3.5	117	11.5	111	98
10	1.5	71	13.0	8.0	151	10	.....	.....	.....	3.5	121	12.0	.....	.....	.....	4.0	100	11.5	3.0	100	11.0	100	95
11	1.5	71	13.0	6.5	123	11	2.0	91	14.0	2.5	86	12.0	5.0	109	12.0	4.0	100	12.0	3.0	100	13.0	97	99
12	.....	.....	.....	6.0	113	13	2.0	91	14.5	2.5	86	14.0	6.0	130	14.0	5.0	125	12.0	2.5	83	11.0	105	103
Av.	2.1	100	12.2	5.3	100	11.7	2.2	100	13.8	2.9	100	12.8	4.6	100	13.6	4.0	100	12.2	3.0	100	11.5	100	12.5
Forecast	3.6	(WNW)	13.0	5.8	(WNW)	12.0	2.0	(WNW)	14.0	2.4	(WNW)	15.0	3.0	(WNW)	15.0	2.6	(WNW)	14.0	3.8	(WNW)	14.0	.....	14.0

1,700 miles WNW of San Diego. The wind blew at 35 knots over a fetch of 800 miles. This situation persisted without much change from May 1 to May 3, 1944. Applying the forecasting method,[14] one obtains 6-foot breakers of 15.5-second period near the Scripps Institution pier at the time the photograph was taken. The height and period, computed from the photograph according to the method given in the surf forecasting manual,[15] equal 7 feet and 14 seconds, respectively. The breaker characteristics computed from weather maps agree with those obtained from the photograph, and the refraction diagram for 14-second waves from the WNW (Fig. 8), should be applicable.

A comparison between Figure 8 and Plate VII shows good agreement between the orientation and spacing of computed and photographed crests. To

[14] "Wind Waves and Swell, Principles in Forecasting," *Hydrographic Office, U.S. Navy, Misc. 11275* (1944). Cf. ftn. 2.

[15] Pp. 42–52 of ftn. 2.

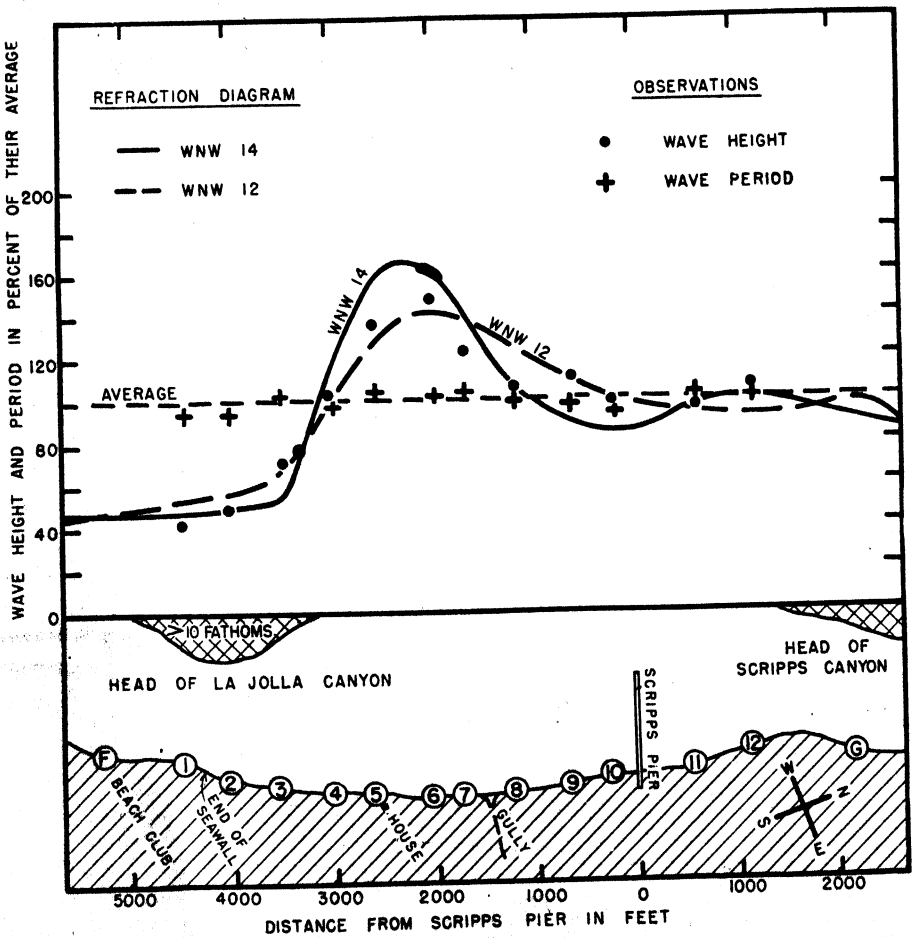

FIG. 16.—Observed versus computed changes in wave height along the beach (see text)

facilitate the comparison, some of the crests have been transferred from Figure 8 to Plate VII, where they are shown by the inked lines. Near the left edge of Plate VII the photographed crests lag, indicating a shallow bank. On a photograph taken 5 months later a similar lag occurred about 500 feet farther south. This evidence indicates the existence of a shifting sand ridge. The depth contours on Figure 8, which are based on soundings taken prior to 1928, cannot take into account such temporary features. It is quite possible that the discrepancy between observed and computed wave heights near Station 7 is also related to this shifting sand bank.

CROSSING OF WAVE TRAINS NORTH OF SCRIPPS CANYON

At the time the refraction diagrams (Figs. 8–12) were drawn, it was expected that they could give only the basic features regarding the positions of crests and the variations in wave height. Accordingly, small-scale features in the bottom topography and sharp bends in the crests or orthogonals were smoothed out, and the smoothed crest was used as a basis for the construction of the next crest. This crest was also smoothed, when necessary, and this smoothing process was continued from crest to crest as the waves were carried toward shore. In this manner a cumulative error is introduced into the refraction diagram.

If these diagrams are drawn with greater precision, and without smoothing, it is found that the steep walls on the northern side of Scripps Canyon cause a section of the crest to be bent through 90°. This section will travel north along the coast and intersect the regular wave train in a crisscross pattern.

The crisscross pattern is illustrated in Figure 17, which shows a large-scale graph of the refraction of 14-second waves from the WNW over Scripps Canyon between Stations 12 and J. The outermost crest has been transferred from Figure 8. The remaining crests were carried forward with a crest interval of one-eighth the wave period and without smoothing, but only one out of eight computed crest positions is shown on the figure (crest interval = one period). The actual construction of the refraction diagram was carried out on a detailed chart, but only two depth contours are shown on the figure.

It should be kept in mind that the shading in Figure 17 designates bottom topography and not convergence or divergence. In order to picture the degree of divergence or convergence, the *width* of the crest lines has been drawn proportional to the refraction factor K according to the scale in the legend. Wherever the refracted crests intersect the main wave train, a hump on the sea surface is formed, the height of which equals the sum of the height of the two intersecting waves.

Plate VIII, *A*, shows a photograph of this wave pattern, taken from the top of the cliffs at a position indicated on Figure 17. The wave crests in the photograph have been slightly retouched to make them appear more clearly in print. The pattern on the photograph resembles closely the pattern on the refraction diagram.

DISCUSSION AND CONCLUSIONS

METEOROLOGY AND WAVE REFRACTION

In order to draw a refraction diagram, it is necessary to know (1) the wave period and direction in deep water and (2) the bottom topography between deep water and the shoreline. The determination of typical wave periods and deep-water directions for a given locality will

be discussed first, and the bottom topography and its effect will be taken up in the following section.

In the experience of the authors, existing observational data on wave characteristics are almost always inadequate for determining typical deep-water wave directions and periods for a given locality. Data on deep-water wave direction are usually lacking entirely, because they require the services of a boat or a plane. Wave period, since it remains constant, can be measured from shore, but the methods of measuring have been for the most part inconsistent and inadequate.

A better procedure is to obtain the wave characteristics from a knowledge of meteorological conditions and the relationships between wind, sea, and swell.[16] Weather maps are examined to determine all wind areas where waves affecting the locality of interest could be generated. The characteristics of the waves upon reaching the given locality are then computed by the relationships mentioned.

This method has been employed in a study for the United States Navy of the effect of wave conditions upon seaplanes. Meteorologists at the Scripps Institution carried out a day-by-day "forecast" for specified localities on the basis of historical weather maps,[17] the forecasts extending over 3 years. This method required considerable work and highly trained personnel. Fortunately, only a small number of localities have to be considered, since offshore conditions are sufficiently uniform to make the computed wave conditions applicable over long stretches of coastline.

[16] See ftn. 7.

[17] A complete set of Northern Hemisphere weather maps, dating back to 1929, were prepared during the war under the auspices of the Joint Meteorological Committee of the Army, Navy, and Weather Bureau.

The procedure described above consists essentially of computing wave characteristics from *synoptic* weather data and then applying statistical methods to the computed wave data for deriving *climatological* wave conditions. In this report a somewhat less accurate procedure has been followed by making use of average wind data obtained from climatological charts for the different seasons. An exceptionally complete set of wave observations at La Jolla, dating back to 1938, has made it possible to check the soundness of this procedure. Even so, it has been necessary to depend entirely upon local observations for information regarding the southern swell which originates in an area of the South Pacific for which weather data are lacking.

BOTTOM TOPOGRAPHY AND WAVE REFRACTION

Detailed hydrographic charts are required for an accurate construction of refraction diagrams. However, the main features in the refraction pattern can be associated with certain types of underwater topography without the need of elaborate computations (Figs. 1–6 and Pls. I–V). Quite generally it can be shown that the wave refraction pattern for any bottom topography follows a single law, according to which **waves change direction and are bent in such a manner that they *tend* to assume the shape of the depth contours.** Thus the wave crests on an aerial photograph draw a quasi-depth chart. This circumstance has found application during the war in determining underwater topography over enemy beaches.

Associated with the distortion of the waves are local convergences and divergences of wave energy. It is important to make a distinction between an area of *divergence* and what might be called a

complete wave *shadow zone.* Waves behind a headland or to the lee of an island are usually very low because of the divergence of wave energy, yet the protected region does not lie in a wave shadow zone. Indeed, sensitive instruments have shown that waves can reach any section of a coastline, no matter how well protected, unless that section is completely cut off from communication with the open sea. By the time the waves have turned through a large angle, they are usually too low to be noticeable to the naked eye, but in some instances local convergences in so-called protected zones have led to surprising results.[18]

THE CONCEPT OF RELATIVE DEPTH

It should be emphasized that the effect of bottom features upon waves depends not upon the absolute depth of water but upon the *relative depth,* that is, the ratio of depth over deep-water wave length (h/L_0). Let h and L_0 be given in feet, and the wave period T in seconds. Setting $g = 32.2$ ft/sec^2 in Appendix equations (A5*b*) and (A6*a*), one obtains approximately $L_0 = 5\ T^2$, and

$$\text{Relative depth}: \quad \frac{h}{L_0} = \frac{h}{5T^2}. \qquad (6)$$

By definition shallow water has a relative depth of less than one-half. For a long 14-second swell shallow water starts therefore at a depth of 500 feet; for a short 7-second wind wave, at 125 feet. Since modification of waves occurs only in shallow water, a shoal at an average depth of 500 feet would affect the swell in a manner similar to that in which a shoal at 125 feet would affect the short-period sea. An interesting application of this principle is the *separation* of a short-period sea from a long-period swell by special types of bottom features. The bottom topography itself can be compared to an enormous lens, separating waves of different frequencies much as a lens breaks up light into the component colors. A famous illustration of this feature is provided by the surf at Waikiki beach in the Hawaiian Islands, which usually results from a NE swell refracting around Nakapuu Point and Diamond Head. Out at sea a short NE chop will usually be present in addition to the swell, but the chop is too short to be refracted effectively and thus passes on without bending into the beach.

Since ordinary waves rarely have periods much longer than 14 seconds, the effect of underwater bathymetry on these waves is virtually limited to the continental shelf. But sea and swell are not the only waves of the sea refracted by the ocean bottom. Fault movements associated with submarine earthquakes generate very long waves, "tsunamis," popularly called tidal waves (although they are not related to tides), and these have periods ranging between 10 minutes and an hour. In accordance with equation (6), these waves are always in shallow water, even over the deepest portion of the oceans. Preliminary results of an investigation by Shepard and others[19] show a good correlation between the variations

[18] An interesting application of the effect of refraction on wave height was made in connection with the liberation of the Philippine Islands. During the planning stage refraction diagrams were prepared for waves entering the Lingayen Gulf from the north, and they showed a convergence for the landing beaches, but a divergence east of the town of Lingayen. Consequently, an alternate plan was made to provide for the contingency of high waves from a typhoon. Shortly after the initial landings, high breakers prevented unloading at the main beachhead, but the breakers remained low to the east of Lingayen as indicated on the refraction diagram. A second beachhead was established according to the alternate plan, and the unloading of supplies continued without interruption.

[19] G. A. MacDonald, F. P. Shepard, and D. C. Cox, "The Tsunami of April 1, 1946, in the Hawaiian Islands (Preliminary Report)," *Pacific Sci.*, Vol. I No. 1 (January, 1947), pp. 21–37.

in the height of the April, 1946, tsunami in the Hawaiian Islands and the offshore topography and promise an interesting application of the principles discussed in this paper. A similar effect, but of even larger order of magnitude, must result from the refraction of tsunamis by large-scale bathymetric features, such as ocean basins and troughs. This might explain the curious fact that waves of the April, 1946, tsunami were recorded about twice as high at South American stations as they were at stations along California, although they had originated near the Aleutian trough and traveled more than twice as far to the South American stations.

WAVE REFRACTION AND BEACH EROSION

Krumbein[20] in his study of Half Moon Bay, California, was probably the first to consider the effect of wave refraction on sediment transport. He states that a refraction diagram indicated at least the order of change in wave energy along the coast. In the present paper it is shown that refraction diagrams also give this change quantitatively. This conclusion is of particular significance because the unusually complex bottom topography off La Jolla provides a critical test of the application of refraction theory to ocean waves. It also appears that wave refraction can be depended upon to give not only the simple main features regarding wave height and direction but also more complicated secondary effects, such as the crisscross pattern north of Scripps Canyon.

Refraction diagrams, if properly used, can become a powerful tool in the study of beach erosion. Although several processes are involved, two must be of particular importance in the effect of refraction upon beach erosion: (1) transport of sediments by *advection* associated with the secondary current system set up by the waves and (2) transport of sediments by *diffusion* associated with the turbulence set up by the breakers. The former process depends critically upon the *direction* of the waves, the latter upon the variation in *height* of the waves. The two processes depend, therefore, upon the two elements of the refraction pattern.

The term "advection" has been borrowed from meteorology or oceanography, where it denotes a transport due to a current flowing from a region of high concentration to a region of low concentration. Mathematically the effects of diffusion and convection are both contained in the vector equation

$$\frac{\partial s}{\partial t} = \nabla^2 \cdot \mathbf{A} - \nabla \cdot s\mathbf{V}, \qquad (7)$$

where s is any concentration, $\mathbf{A}$ the vector coefficient of kinematic eddy diffusivity, and $\mathbf{V}$, the vector velocity. Equation (7) states that the local time change of any concentration equals the effect of diffusion minus the effect of advection. For further discussion refer to chapter v of Sverdrup *et al.*[21]

Of the currents set up by the waves, the most important are probably longshore currents, or littoral currents, and rip currents.[22] The longshore currents are set up by the transport of waves breaking at an angle with the shoreline. They are directed, therefore, *with* the longshore component of the wave motion—for example, from left to right in Figure 1. In the case of high waves and breakers forming an angle of more than 10° with

[20] "Shore Processes and Beach Characteristics," *Technical Memorandum No. 3, Beach Erosion Board* (May, 1944).

[21] H. U. Sverdrup, M. W. Johnson, and R. H. Fleming, *The Oceans, Their Physics, Chemistry and General Biology* (New York: Prentice-Hall, 1942).

[22] F. P. Shepard, K. O. Emery, and E. C. La Fond, "Rip Currents: A Process of Geological Importance," *Jour. Geol.*, Vol. XLIX, No. 4 (May–June, 1941), pp. 337–69.

the coastline, longshore currents may attain velocities of several knots. If it were not for the fact that breaker angles are greatly reduced by refraction, longshore currents would attain such high velocities that the rate of beach erosion would be many times its actual value.

The refraction pattern for submarine canyons causes the longshore currents to diverge from the center of the canyon toward both sides, leading to a removal of sediment from the canyon head. In the case of submarine ridges, the longshore currents and the sediments converge upon the ridge. Submarine canyons and ridges are associated, therefore, with a refraction pattern which sets up a current system that tends to perpetuate their existence.

Rip currents are directed from shore out to sea. Along a straight coastline they are most pronounced when the waves come straight toward shore. Usually the location of rip currents is closely associated with the refraction pattern. In the case of submarine canyons, rip currents tend to form at the zones of convergence at either side of the canyon, although frequently they turn and flow directly over the canyon just outside the breaker zone. In the case of ridges, rip currents tend to form in the convergence above the ridges. Exceptions to all these rules are, however, frequently found. The effectiveness of rip currents in transporting sand can be gauged by the fact that they have been observed to erode a channel several feet deep within a short time interval.[23]

Transportation of sediments by *diffusion* must result from the variation of wave heights along the beach. Wave heights are low opposite the mouth of the submarine canyons and high on either side. Since high breakers cause a heavy degree of turbulence, and low breakers only little turbulence, sand may be kept in suspension on the sides of the canyons while it has a chance to settle over the canyon. In the case of submarine ridges similar reasoning leads to the conclusion that the ridges tend to be eliminated by diffusion. Thus the effect of diffusion is opposite to the effect of advection. Since it is known that the heads of submarine canyons are kept fairly clean of sediments[24] and that ridges may exist for long times, it appears that the advective processes must balance the effects of diffusion and slumping.

Tidal currents are rarely of much importance along open coastlines exposed to severe wave action. The change in tide level may, however, be important, since it causes a back-and-forth shift of the surf zone, which in turn will have an effect upon the location of longshore currents, rip currents, and the zones of maximum turbulence. The rise and fall of the mean sea-level makes it impossible for any beach ever to reach a state of complete equilibrium.

It should be emphasized that the manner in which sediment is transported by wave action is not properly understood and that many exceptions have been found to the foregoing statements. Conditions, as they exist along a beach at any given time, approach a sensitive balance between erosion and deposition. The balance is profoundly disturbed by any changes in the wave and weather conditions or by the building of jetties and breakwaters. For similar reasons the manner in which the sands along a beach are shifted back and forth by wave action must depend greatly upon the refraction pattern. The resulting sediment transport may cause changes in depth which in turn affect the refraction pat-

[23] *Ibid.*, pp. 350–52.

[24] Shepard and Emery, pp. 94–103 of ftn. 13.

tern (Pl. VIII, *B*). This interplay between waves and bottom makes the prediction of changes in the bottom profiles extremely difficult. But by analyzing the effects of wave action from the point of view of fluid mechanics, making use also of refraction diagrams, it seems likely that a better understanding of the processes involved in beach erosion will result.

Another point of view we have tried to introduce into this report is the importance of meteorology to studies of beach erosion. A season which is unusual from the point of view of beach conditions must, in the final analysis, be the result of unusual meteorological conditions over the ocean. This does not necessarily mean that local weather conditions have been unseasonal. We have noted, for example, that some of the swell reaching southern California in the summer depends upon weather conditions in the Southern Hemisphere, perhaps 5,000 miles away. The ever changing beaches along the shorelines are a reflection of the infinite number of possible combinations and variations of storms over widely spread areas of the ocean.

APPENDIX

THEORY OF WAVE REFRACTION

Let E denote the mean energy per unit surface area, n the fraction of this energy advancing with the wave crest at velocity C, and s the distance between adjacent orthogonals. Then, if energy is conserved between orthogonals,

$$E_0 n_0 C_0 s_0 = EnC\,s = E_b n_b C_b s_b\,, \quad \text{(A1)}$$

where the subscript "o" refers to deep-water conditions, i.e., water whose depth exceeds one-half the wave length, and the subscript b to the breaker point. The expression without subscripts holds generally in shallow water, intermediary between deep water and the breaking-point.

In deep water the wave energy is proportional to the *square* of the wave height H_0 according to the equation

$$E_0 = \tfrac{1}{8}\rho g H_0^2\,, \quad \text{(A2}a\text{)}$$

where ρ is density and g the acceleration due to gravity. The same relationship holds with good approximation in shallow water:

$$E = \tfrac{1}{8}\rho g H^2\,. \quad \text{(A2}b\text{)}$$

However, just outside the breaking-point, the shape of the waves approaches that of "solitary waves," with sharp isolated crests, separated by long flat troughs. For solitary waves the wave energy is proportional to the cube of the wave height, according to the equation

$$E_b = \frac{\rho g}{L_b}\left(\frac{4h_b H_b}{3}\right)^{3/2}. \quad \text{(A2}c\text{)}$$

Combining the first two expressions of equations (A1) and (A2), we get the following general relationship for the wave height in shallow water:

$$\frac{H}{H_0} = \left(\frac{n}{n_0}\frac{C}{C_0}\frac{s}{s_0}\right)^{-1/2}. \quad \text{(A3)}$$

In deep water only half the energy, the potential energy, advances with the wave form:[25]

$$n_0 = \tfrac{1}{2}\,. \quad \text{(A4}a\text{)}$$

In the breaker zone both potential and kinetic energy travel with the waves,

$$n_b = 1\,, \quad \text{(A4}b\text{)}$$

while for shallow water in general

$$n = \frac{1}{2}\left[1 + \frac{\frac{4\pi h}{L}}{\sinh\left(\frac{4\pi h}{L}\right)}\right]. \quad \text{(A4}c\text{)}$$

Equation (A4*c*) reduces properly to equations (A4*a*) or (A4*b*), depending upon whether the ratio h/L has a large or small value.

The velocity of sinusoidal irrational waves of low amplitude is given by the classical Stokes's equation

$$C^2 = \frac{gL}{2\pi}\tanh\frac{2\pi h}{L}\,, \quad \text{(A5}a\text{)}$$

[25] H. U. Sverdrup and W. H. Munk, "Wind, Sea and Swell: Theory of Relations for Forecasting," *Hydrographic Office, Technical Report in Oceanography No. 1* (in press).

which for deep water reduces to the form

$$C_0^2 = \frac{gL_0}{2\pi}. \qquad \text{(A5}b\text{)}$$

Since wave period T, defined as the ratio of wave length over wave velocity, remains constant

$$T = \frac{L_0}{C_0} = \frac{L}{C} = \frac{L_b}{C_b} \qquad \text{(A6}a\text{)}$$

or

$$\frac{C}{C_0} = \frac{L}{L_0}. \qquad \text{(A6}b\text{)}$$

Dividing (A5a) by (A5b) and making use of (A6b), we get

$$\frac{C}{C_0} = \frac{L}{L_0} = \tanh \frac{2\pi h}{L}. \qquad \text{(A7)}$$

Equation (A7) can also be written

$$\frac{L}{L_0} = \tanh \left[2\pi \frac{h}{L_0} \left(\frac{L}{L_0} \right)^{-1} \right], \qquad \text{(A7}'\text{)}$$

which is an equation of the form

$$\frac{L}{L_0} = \text{function} \left(\frac{h}{L_0} \right). \qquad \text{(A8)}$$

The ratio h/L_0 is termed "relative depth." It follows from equations (A7) and (A8) that the ratio C/C_0 is a function of the relative depth only. In a similar manner it can be shown that the ratio n/n_0 is a function of the relative depth.

Thus equation (A3), which can be written

$$\frac{H}{H_0} = \left(\frac{n}{n_0} \frac{C}{C_0} \right)^{-1/2} \sqrt{\frac{s_0}{s}}, \qquad \text{(A9)}$$

consists of two parts: the first part, which depends upon the relative depth and does not contribute toward a *variation* in wave height along any fixed contour, and the second part, the refraction factor, according to which any variation in wave height along a beach outside the breaker zone, must be proportional to the square root of s_0/s.

To find the corresponding expression for the breakers, combine the first and third term of equations (A1) and (A2)

$$\frac{C_0}{2} \tfrac{1}{8} H_0^2 s_0 = \frac{C_b}{L_b} \left(\frac{4 h_b H_b}{3} \right)^{3/2} s_b,$$

where use has been made of equations (A4a) and (A4b). In view of (A6a), this can be written

$$h_b H_b = \tfrac{3}{4} \left(\tfrac{1}{16} \right)^{2/3} (H_0^2 L_0)^{2/3} \left(\frac{s_0}{s_b} \right)^{2/3}. \qquad \text{(A10)}$$

According to theory substantiated by experiments with solitary waves,

$$h_b = 1.28 H_b, \qquad \text{(A11)}$$

and equation (A10) becomes

$$\frac{H_b}{H_0} = \frac{1}{3.3 \sqrt[3]{\frac{H_0}{L_0}}} \sqrt[3]{\frac{s_0}{s}}. \qquad \text{(A12)}$$

Any *variation* in breaker height must be proportional to the cube root of s_0/s. The ratio H_0/L_0 is called the wave steepness in deep water. Equation (A12) and the foregoing discussion hold strictly only for waves of steepness H_0/L_0 less than 1 per cent (i.e., for swell) but is approximately true for all but the steepest storm waves.

The general equation (A9) for wave heights in shallow water neglecting refraction was first derived by Green in 1839 and extended by O'Brien and Mason[26] to include the refraction effect. The existence of solitary waves was first noticed by Russell, and their theory was developed by Boussinesque in his study of tide waves and related phenomena. The application of the solitary-wave theory to surf problems and the derivation of equation (A12) are believed to be new.

ACKNOWLEDGMENTS.—We are indebted to Dr. H. U. Sverdrup for the active role he has played in developing the theory and application of wave refraction. Dr. F. P. Shepard and Mr. D. Leipper have made many helpful suggestions. The exacting job of constructing the refraction diagrams in Figures 8–12 was undertaken by Sgt. R. Jentoft, AC, while on duty at the Scripps Institution. Mr. D. Pritchard has drawn the complex refraction diagram shown in Figure 17.

[26] Pp. 37–39 of ftn. 5.

PLATE I

A, Long waves from the southern hemisphere approaching the coast north of Oceanside, California. Compare refraction pattern with schematic drawing in Fig. 1. (Official photograph, United States Navy.)

B, Refraction of waves over ridge along the northern shore of Oahu, T.H. (cf. Fig. 4). Over the ridge waves lag behind and are higher than on either side. (Official photograph, United States Army.)

PLATE II

Refraction of waves over submarine canyon at Redondo, California (cf. Fig. 3). Waves bend sharply forward over canyon. Waves are higher just outside the canyon than over the canyon, and the surf zone is narrow opposite the mouth of the canyon. (Photograph by Spence Air Photos, Los Angeles.)

PLATE III

Refraction of waves over the coral reef at Guam. Wave crests move ahead over the channels and lag behind over the shoals. Wave heights are low over channels and large over shoals. Compare with schematic drawings in Figs. 3 and 4. (Official photograph, United States Navy.)

PLATE IV

A, Refraction of waves around Point Loma, San Diego, California. Solid orthogonal shows main refraction around the point; dashed orthogonal designates secondary refraction around a shoal directly off the end of Point Loma. Notice variation in width of surf zone. (Official photograph, United States Navy.)

B, Waves turn through 180° around the hook of Cape Cod but, in doing so, lose most of their height because of the very large divergence. Notice variation in width of surf zone. (Official photograph, United States Navy.)

PLATE V

Waves bend toward the famous Waikiki Beach on Oahu, T.H. (Official photograph, United States Navy.)

PLATE VI

Photograph of a three-dimensional model of the bottom topography off La Jolla, California. Two submarine canyons, La Jolla Canyon and its steep-walled tributary, Scripps Canyon, reach within a thousand feet from shore.

PLATE VII

Observed versus computed orientation and spacing of wave crests. The computed positions (*inked lines*) are parallel to the positions of the crests on the photograph, except for a small discrepancy on the left side of the figure (see text).

PLATE VIII

A, Photograph of waves north of Scripps Canyon shows the presence of the crisscross pattern which appears in the refraction diagram (Fig. 17). Peaks occur where the refracted crests intersect the crests of the main wave train.

B, Refraction of waves at boat basin, Oceanside, California. Longshore currents set up by waves have widened the beach on both sides of jetties (see also Pl. II). This in turn has caused refraction of waves near jetties. The resulting refraction pattern tends to further strengthen the convergence of longshore currents upon the jetties and may lead to accelerated deposition.

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PLATE I

PLATE II

PLATE III

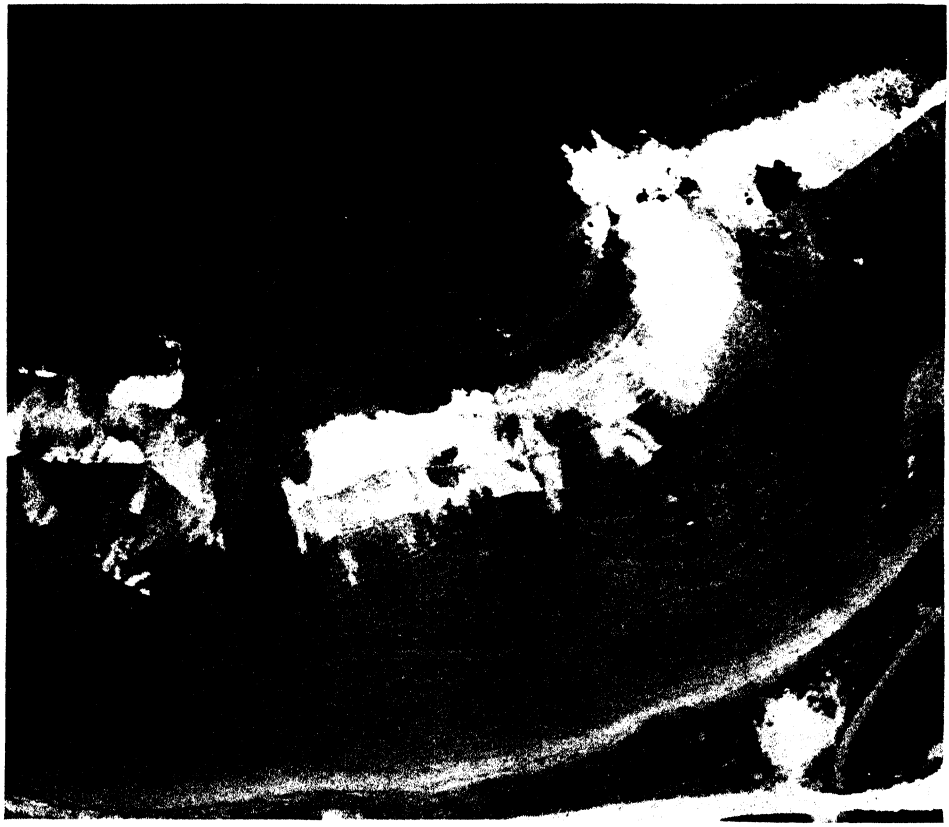
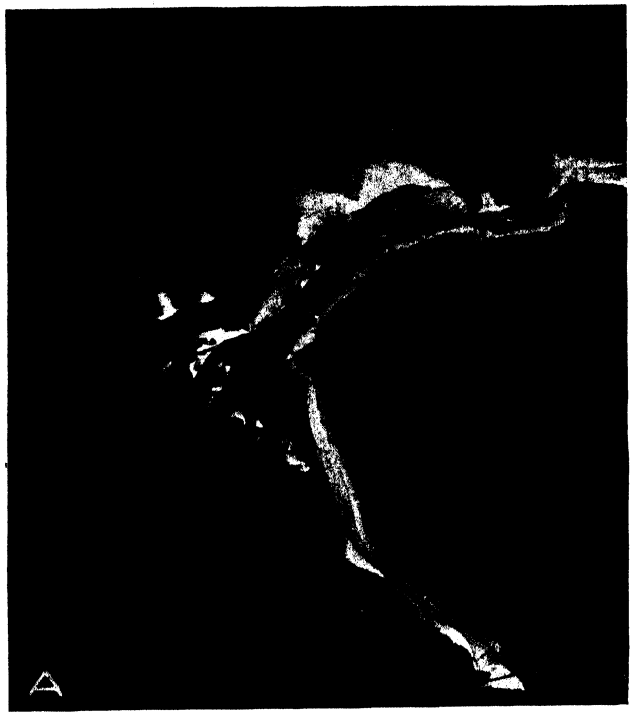


PLATE IV



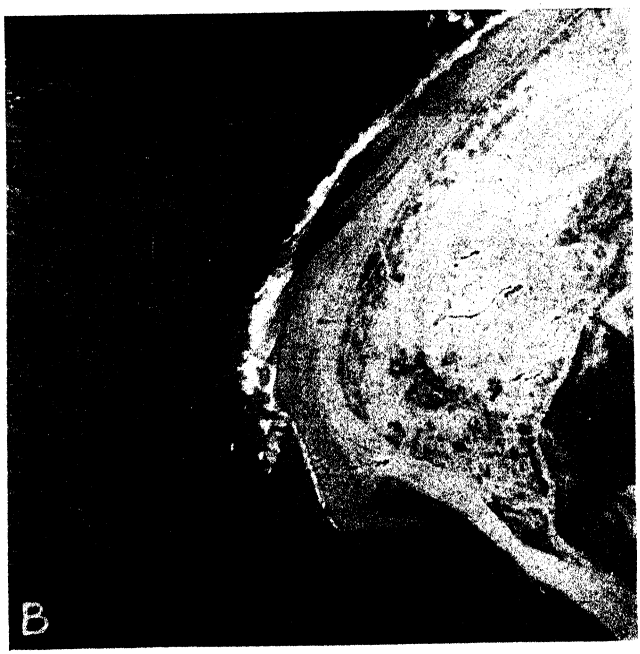


PLATE V

PLATE VI

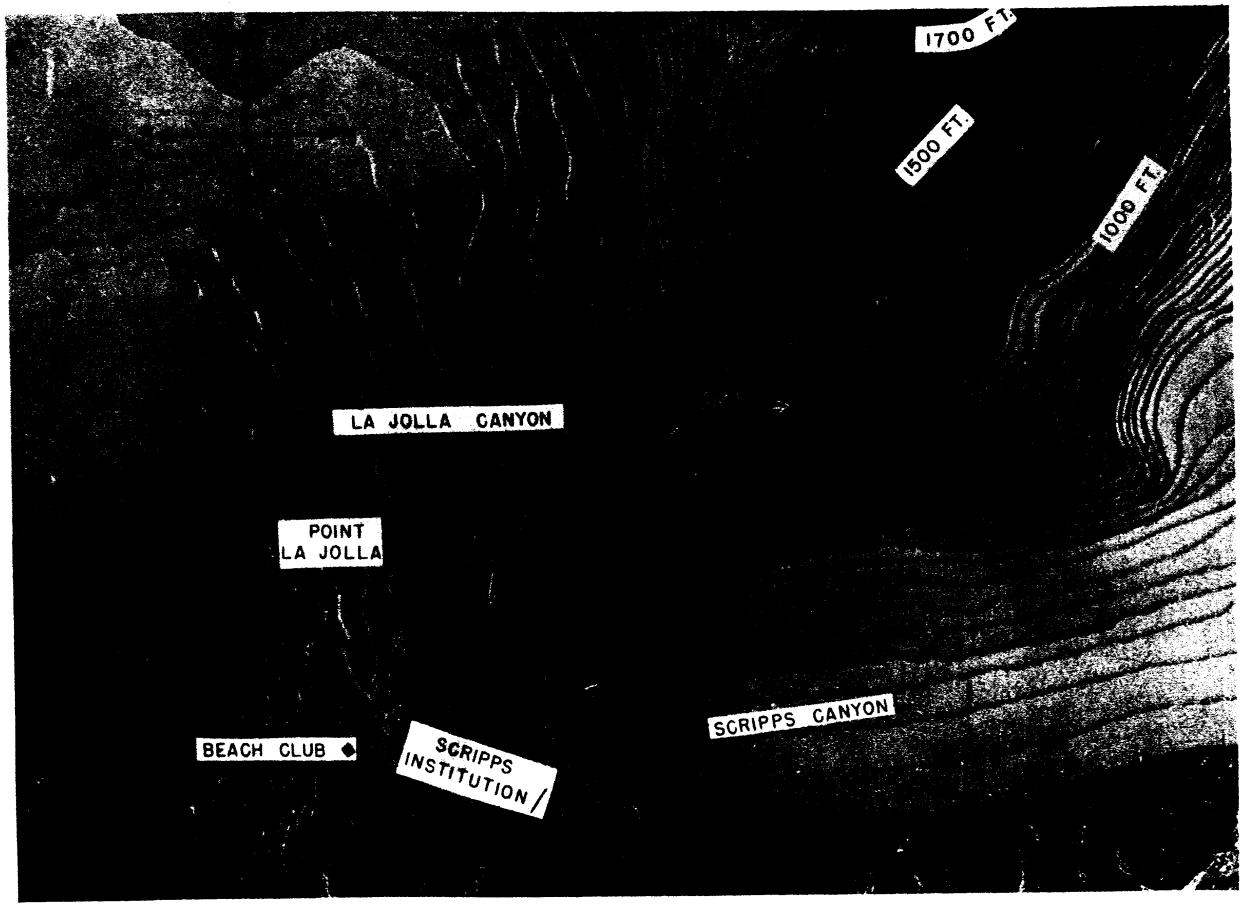
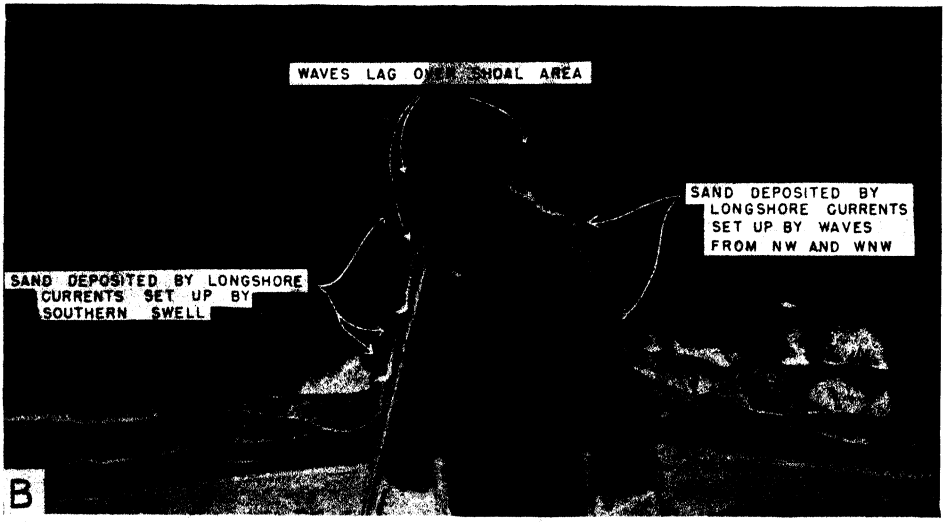

1700 FT.
1500 FT.
1000 FT.
LA JOLLA CANYON
POINT LA JOLLA
BEACH CLUB
SCRIPPS INSTITUTION
SCRIPPS CANYON

PLATE VII

PLATE VIII

A


WAVES LAG O SHOAL AREA
SAND DEPOSITED BY LONGSHORE CURRENTS SET UP BY WAVES FROM NW AND WNW
SAND DEPOSITED BY LONGSHORE CURRENTS SET UP BY SOUTHERN SWELL
B

THE SYSTEM $NaAlSiO_4$-$Ca_2Al_2SiO_7$

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University of Chicago

ABSTRACT

The high-temperature phase equilibrium relations in the system $NaAlSiO_4$-$Ca_2Al_2SiO_7$ were investigated experimentally by the quenching method. The behavior of this system was found to differ from that of a true binary system in a number of important respects, and it is concluded that the system is actually quaternary. An illustration is thus provided of the extent to which equilibrium relations can be rendered complex, even in such comparatively simple melts, by the great tendency of certain minerals to form complex solid solutions.

INTRODUCTION

This investigation is a study of the high-temperature equilibrium relationships, at atmospheric pressure, of mixtures of the two silicates, sodium aluminum orthosilicate ($NaAlSiO_4$) and the calcium aluminum silicate, $Ca_2Al_2SiO_7$. These compounds are the principal molecules[1] of the natural minerals nepheline and gehlenite, respectively. Both nepheline and gehlenite are feldspathoids, important constituents of the alkaline series of igneous rocks. Nepheline is the most characteristic mineral of these rocks, while melilites, of which mineral group gehlenite is an end-member, are found in alkaline rocks rich in lime. Nepheline and melilite are found in association in some igneous rocks. Gehlenite also often occurs as a metamorphic mineral.

The compositions of all possible mixtures of the compounds $NaAlSiO_4$ and $Ca_2Al_2SiO_7$ may be expressed in terms of the four oxides, soda, lime, alumina, and silica. The system under consideration forms an internal linear join in the composition tetrahedron which has these four oxides as its apexes. The two compounds are represented by points at the ends of this join, lying on faces of the tetrahedron.

The properties of the two individual silicates were not determined in this investigation, as this information was available from the findings of other workers. Eleven different mixtures of the two components were prepared from the oxides. These were studied thermally by means of the quenching method, in which a small charge is allowed to come to equilibrium at a known temperature and then chilled rapidly by being dropped into mercury. It is then examined microscopically to determine the phases present at equilibrium. About 160 "runs" of this kind were made.

PROPERTIES OF THE CRYSTALLINE PHASES

CARNEGIEITE

Carnegieite, the high-temperature form of $NaAlSiO_4$, exists in three forms, none of which occurs in nature. It melts at a temperature of 1526° C.[2] and is isometric at high temperatures, altering metastably to a triclinic form upon cool-

[1] In spite of our present knowledge of crystals as atomic lattices and of the laws controlling the proxying of one atom for another, the concept of molecules in solid solution in crystal phases of variable composition is still a useful one, especially in studies of phase equilibriums.

[2] N. L. Bowen, "The Binary System: $Na_2Al_2Si_2O_8$ (Nephelite, Carnegieite)-$CaAl_2Si_2O_8$ (Anorthite)," *Amer. Jour. Sci.*, Vol. XXXIII (4th ser., 1912), pp. 554 ff.

ing below 690° C.[3] A second inversion occurs at a temperature of 226.5° C., accompanied by a marked change in birefringence. The indices of refraction of low-temperature carnegieite are given by Bowen[4] $\gamma = 1.514$, $\beta = 1.514-$, $\alpha = 1.509$. Carnegieite usually lacks crystal form, occurring in small globular masses, and typically has an intricate polysynthetic twinning.

The carnegieite obtained in the present study had typical globular form, and much of it exhibited characteristic twinning. Its indices were always significantly higher than those of pure carnegietite, however, and thus indicated the presence of solid solution. This also resulted in a decrease in birefringence to the extent that separate indices could rarely be distinguished. A trend toward higher carnegieite indices was observed in mixtures successively richer in $Ca_2Al_2SiO_7$. For example, the average index for carnegieite obtained in runs near the liquidus temperatures increased from 1.516 in the 90 per cent $NaAlSiO_4$ mixture to 1.522 in the 70 per cent $NaAlSiO_4$ mixture. Likewise, the average index was higher in carnegieite which was formed at lower temperatures. Thus, in the case of the 80 per cent $NaAlSiO_4$ mixture, the index changed from 1.520 in carnegieite formed at 1399° C. to 1.523 in that formed at 1187° C. The trend toward higher indices of carnegieite in mixtures successively richer in lime appeared to reverse itself after the 30 per cent $Ca_2Al_2SiO_7$ mixture was passed. These variations in the indices of the carnegieite in this system are significant as indications of the extent to which the composition of this phase was modified by solid solution.

NEPHELINE

Pure artificial nepheline inverts to carnegieite at a temperature of 1254 ±5° C.[5] However, Bowen[6] found that solid solution of anorthite raised the inversion point of nepheline as much as 100°, while albite in solid solution has been shown[7] to raise the inversion temperature some 30°. In the present study this transformation took place over a range of temperatures extending from about 50° above to 100° below the normal inversion point.

Nepheline is hexagonal and crystallizes in melts in short prismatic crystals with basal pinacoid. Bowen[8] gives the indices as $\epsilon = 1.533$ and $\omega = 1.537$. He found that solid solution of anorthite raised the value of ϵ to as high as 1.539, ω remaining constant. Solid solution of albite in nepheline has been found[9] to lower the indices of the latter somewhat. The nepheline obtained in the present study had unusually high indices, which increased from an average value of 1.540 in runs made with the 90 per cent $NaAlSiO_4$ mixture to as high as 1.552 in the 65 per cent $NaAlSiO_4$ mixture. The birefringence usually was very low, and this made it impossible to determine optic sign except in a few cases, in which it was negative. Two distinct indices could not be measured, and the sign of elongation of the crystals was only rarely ascertainable.

[3] N. L. Bowen and J. W. Greig, "The Crystalline Modifications of $NaAlSiO_4$," *Amer. Jour. Sci.*, Vol. X (5th ser., 1925), pp. 208–10.

[4] P. 564 of ftn. 2.

[5] J. W. Greig and Tom F. W. Barth, "The System $Na_2O \cdot Al_2O_3 \cdot 2SiO_2$ (Nephelite, Carnegieite)-$Na_2O \cdot Al_2O_3 \cdot 6SiO_2$ (Albite)," *Amer. Jour. Sci.*, Vol. XXXV-A (5th ser., 1938), p. 108.

[6] P. 571 of ftn. 2.

[7] P. 109 of ftn. 5.

[8] P. 566 of ftn. 2.

[9] W. R. Foster, "The System $NaAlSi_3O_8$-$CaSiO_3$-$NaAlSiO_4$," *Jour. Geol.*, Vol. L (1942), p. 157.

GEHLENITE

Artificial gehlenite melts congruently at a temperature of 1590° C.[10] It is tetragonal, forming short prismatic or tabular crystals with distinct basal cleavage. Its indices of refraction are 1.669(ω) and 1.658(ϵ). The melilite crystals encountered in this investigation usually had the typical crystal habit, but sometimes they assumed shapeless rounded forms, clustered together and often very minute. They differed considerably from pure gehlenite in indices of refraction. Although the indices could be determined with accuracy in only a relatively few runs, sufficient data were obtained to show a progressive decrease in indices with increase in $NaAlSiO_4$ content of the mixture. Thus, in runs made with the 30 per cent $NaAlSiO_4$ mixture the melilite indices were 1.666 ($\pm$.002) and 1.658, while in a run made with the 40 per cent $NaAlSiO_4$ mixture the indices were found to be 1.662 and 1.655. In the 60 per cent $NaAlSiO_4$ mixture, the melilite index ω reached a value as low as 1.651. In mixtures richer in $NaAlSiO_4$ than 60 per cent, the indices were somewhat higher, indicating that the trend toward lower indices is reversed at about this composition. Considering a single mixture, the indices were somewhat lower in melilite formed at lower temperatures. Thus, in runs made with the 40 per cent nepheline mixture, the indices of melilite formed at 1408° C. were 1.662 and 1.655, while those of melilite formed at 1247° C. were 1.656 and 1.650. These differences in optical properties give some indication of the extent of the variations in composition of the "gehlenites" of this system.

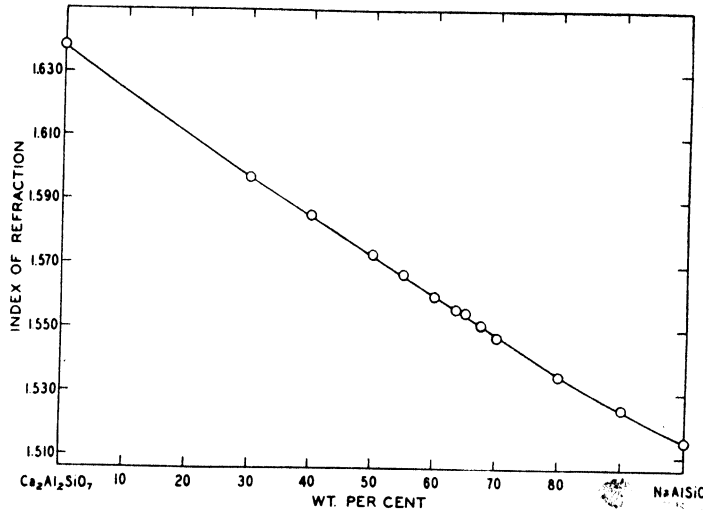

FIG. 1.—Indices of refraction for the glasses of the system $NaAlSiO_4$-$Ca_2Al_2SiO_7$

THE THERMAL DATA

In the thermal investigation eleven mixtures of different composition were prepared and studied. The refractive indices of the glasses of these mixtures are given in the graph of Figure 1. Of the 160 runs which were made, less than

[10] E. F. Osborn and J. F. Schairer, "The Ternary System, Pseudo-wollastonite-Akermanite-Gahlenite," *Amer. Jour. Sci.*, Vol. CCXXXIX (1941), p. 719.

TABLE 1

THERMAL DATA FOR THE SYSTEM $NaAlSiO_4$-$Ca_2Al_2SiO_7$

Weight Per Cent		Initial Condition	Time	Temp. (° C.)	Final Condition
$NaAlSiO_4$	$Ca_2Al_2SiO_7$				
30	70	glass	60 min.	1469	MEL and glass
		glass	35 min.	1473	All glass
40	60	glass	45 min.	1415	MEL and glass
		glass	40 min.	1420	Rare MEL and glass
		glass	60 min.	1424	All glass
50	50	glass	2 hr.	1230	MEL, rare CG, and glass
		devitr.	2 hr.	1281	MEL, rare CG, and glass
		devitr.	2 hr.	1285	MEL and glass
		glass	40 min.	1356	Rare MEL and glass
		glass	40 min.	1360	All glass
55	45	glass	2¼ hr.	1260	MEL, CG, rare NE, and glass
		devitr.	2 hr.	1264	MEL, CG, and glass
		devitr.	2 hr.	1281	MEL, CG, and glass
		devitr.	2¼ hr.	1285	MEL and glass
60	40	devitr.	2 hr.	1281	MEL, CG, NE, and glass
		devitr.	2¼ hr.	1285	MEL and glass
		glass	90 min.	1300	MEL and glass
		glass	45 min.	1304	All glass
63.5	36.5	devitr.	2 hr.	1218	MEL, NE, rare CG, and glass
		devitr.	2½ hr.	1281	MEL, NE, rare CG, and glass
		devitr.	2¼ hr.	1285	MEL, NE, and glass
		devitr.	2 hr.	1289	Very rare MEL and glass
65	35	devitr.	2 hr.	1218	MEL, NE, rare CG, and glass
		devitr.	2 hr.	1277	NE, MEL, rare CG, and glass
		devitr.	2½ hr.	1281	NE, MEL, and glass
		devitr.	2¼ hr.	1285	NE and glass
		glass	1 hr.	1290	Rare NE and glass
67.5	32.5	devitr.	2 hr.	1257	MEL, NE, rare CG, and glass
		devitr.	2 hr.	1274	CG, NE, MEL, and glass
		devitr.	2½ hr.	1278	CG, NE, and glass
		devitr.	2 hr.	1290	CG, NE, and glass
		devitr.	2¼ hr.	1294	Rare CG and glass
		glass	1½ hr.	1298	All glass
70	30	devitr.	2 hr.	1195	NE, MEL, and glass
		devitr.	2½ hr.	1199	NE, MEL, rare CG, and glass
		devitr.	2 hr.	1276	NE, CG, rare MEL, and glass
		devitr.	3 hr.	1280	CG, NE, and glass
		devitr.	1½ hr.	1296	CG, NE, and glass
		devitr.	2 hr.	1300	CG and glass
		glass	75 min.	1321	Rare CG and glass
		glass	40 min.	1325	All glass
80	20	devitr.	2 hr.	1154	NE, MEL, and glass
		devitr.	2 hr.	1171	NE, MEL, rare CG, and glass
		devitr.	½ hr.	1268	CG, NE, rare MEL, and glass
		devitr.	2 hr.	1272	CG, NE, and glass
		devitr.	3½ hr.	1296	CG, NE, and glass
		devitr.	2 hr.	1298	CG and glass
		glass	50 min.	1399	Rare CG and glass
		glass	40 min.	1404	All glass
90	10	devitr.	2 hr.	1103	NE, MEL, and glass
		devitr.	2 hr.	1154	NE, MEL, rare CG, and glass
		devitr.	3 hr.	1229	NE, CG, very rare MEL, and glass
		glass	2 hr.	1231	NE, CG, and glass
		devitr.	2 hr.	1274	CG, rare NE, and glass
		devitr.	2½ hr.	1285	CG and glass
		glass	60 min.	1465	Rare CG and glass
		glass	[illegible] min.	1469	All glass

half were sufficient to fix the temperatures of the various transformations. The results of the critical runs are given in Table 1 on page 30. The equilibrium diagram for the system appears as Figure 2.

NONBINARY NATURE OF THE SYSTEM

Even a casual inspection of the equilibrium diagram will suffice to show that this is not a true binary system. As pointed out above, the variation of optical properties indicates that each of the three crystalline phases occurring in the system is actually a solid solution. The equilibrium diagram shows clearly that these solid solutions are not composed of the two binary components alone, but that other substances, not representable in terms of $NaAlSiO_4$ and $Ca_2Al_2SiO_7$, are also present. Perhaps the best evidence for this is the fact that the system lacks a binary eutectic. Nor are any simple solidus curves present, as would be the case were the solid solutions composed of the two components alone. In fact, liquid persists down to the lowest temperatures of observation throughout the system. During crystallization, the second solid phase does not make its appearance at any fixed temperature (eutectic) but begins to crystallize at a different temperature in each mixture. The inversion relations of nepheline and carnegieite likewise attest to the nonbinary nature of the system. Inversion takes place over a wide temperature range in mixtures on both sides of the pseudoeutectic composition. In compositions adjacent to the latter, nepheline is in equilibrium with liquid at higher temperatures than is carnegieite!

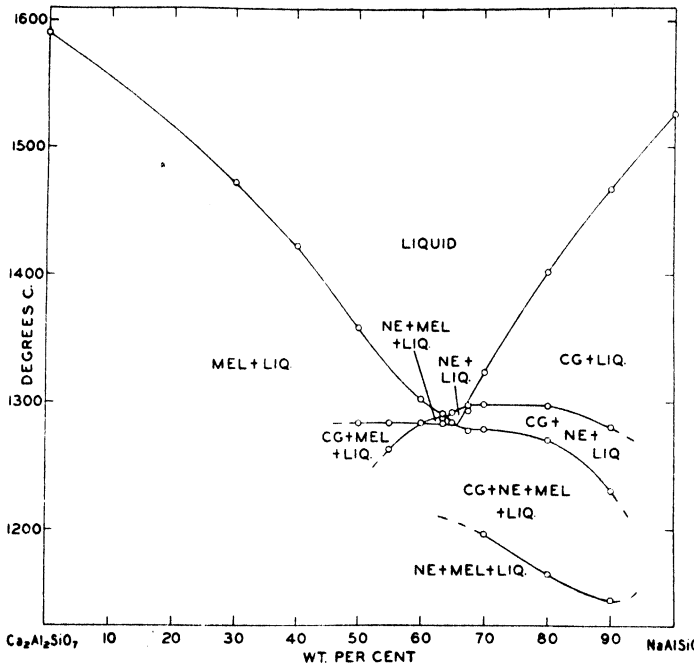

FIG. 2.—Equilibrium diagram for the system $NaAlSiO_4$-$Ca_2Al_2SiO_7$

From a consideration of the phase rule, which, as applied to a condensed system, is $P + F = C + 1$, it is clear that no more than three phases can occur in equilibrium in a binary system, and as many as three only at a fixed temperature. In one field of the equilibrium diagram for the present system four phases coexist, and in several other fields three phases occur in equilibrium. As the system is therefore not binary, the substances entering into the carnegieite, nepheline, and melilite solid solutions cannot be expressed in terms of two components. The components of the system must be considered four—the four constituent oxides present in all compositions. As many as five phases could coexist in equilibrium at an invariant point in such a system. The liquidus curves of the diagram do not then represent univariant equilibriums, but are trivariant, and the pseudoeutectic is actually divariant. The compositions of the phases lie outside the plane of the diagram and hence are not determinable for specific temperatures and compositions from this diagram.

THE LIQUIDUS RELATIONS

The melting-point of pure $Ca_2Al_2SiO_7$ is 1590° C., and that of pure $NaAlSiO_4$ is 1526° C. From these points the liquidus curves slope downward toward each other in the usual way, forming a junction at a temperature of about 1287° C. and a composition of 64 per cent $NaAlSiO_4$. As pointed out above, this is not a true eutectic point. The carnegieite liquidus has a sharp break in slope at its junction with the upper boundary of the carnegieite-nepheline inversion range—the usual effect of inversion. This junction occurs at about 1297° C. and a composition of 67 per cent $NaAlSiO_4$. Each point on these curves represents the temperature at which the first crystals appear as a charge of that composition is cooled, and the temperature where melting is completed as such a charge is heated up. The curves do not, as in a true binary system, give the compositions of the liquids in equilibrium with the separating crystals at the various temperatures. Instead, the composition of each liquid follows a path which is outside the plane of the diagram and is different for each mixture. This is because the compositions of the mix-crystals separating out do not lie in the plane of the diagram. As these crystals form, the liquid in equilibrium with them becomes progressively impoverished in the substances going to make up the crystals, and its composition necessarily leaves the plane of the diagram.

THE SECONDARY-PHASE BOUNDARIES

From the pseudoeutectic point, the secondary-phase boundaries slope downward on both sides, behaving quite unpredictably. Any point on these curves gives the temperature at which the second phase[11] makes its appearance during the cooling of a melt of the indicated composition. Liquid persists below these boundaries, although it slowly diminishes in amount as the temperature is lowered. The curve on the nepheline side of the diagram is quite gentle in slope as it proceeds away from the pseudoeutectic point, but curves rather sharply at about 80 per cent $NaAlSiO_4$, and dips steeply from that point on. The secondary-phase boundary on the gehlenite side of the diagram is nearly horizontal as far from the pseudoeutectic point as it was determined. For a short distance near that point this curve marks the upper limit of nepheline as the secondary phase.

[11] On the nepheline side of the diagram, it is actually the *third* solid phase which begins to form at this boundary, as the carnegieite-to-nepheline inversion has not been completed, and both forms of $NaAlSiO_4$ are present.

From approximately the 40 per cent $Ca_2Al_2SiO_7$ composition toward mixtures richer in that component, it lies above a field in which carnegieite is the secondary phase.

THE NEPHELINE-CARNEGIEITE INVERSION

The inversion relations between carnegieite and nepheline are quite unusual. The inversion takes place over a large range of temperatures. The upper boundary of the inversion range gives the temperature at which the first nepheline appears on cooling a melt and the last traces of nepheline disappear on heating it. Likewise, the lower boundary marks the completion of the carnegieite-to-nepheline inversion and the beginning of the opposite transformation. The upper limit of the inversion range on the $NaAlSiO_4$ side of the diagram lies at approximately 1297° C. from its junction with the liquidus to a composition somewhat richer in $NaAlSiO_4$ than 80 per cent, where it falls rather rapidly toward the normal inversion point at 1254° C. The lower limit of the carnegieite-nepheline transformation was found to be, at the 90 per cent $NaAlSiO_4$ composition, somewhat lower than 1150° C., or 100° below the normal inversion temperature. Between this composition and that of pure $NaAlSiO_4$ this boundary must slope upward toward the temperature 1254° C. In the other direction this curve also slopes upward, probably continuing to rise over into the melilite side of the diagram.

Strangely enough, nepheline is stable at higher temperatures than carnegieite for a small range of compositions close to the pseudoeutectic point. It is difficult to find an explanation for this behavior solely on the basis of the results of the present study, but it is hoped that studies of adjoining mixtures will shed light on the problem. From approximately the 40 per cent $Ca_2Al_2SiO_7$ composition toward compositions richer in that component, carnegieite is again stable at higher temperatures than is nepheline. The upper limit of inversion falls rapidly in this direction, at least where it could be definitely determined. It was useless to attempt to fix the various boundaries farther toward the $Ca_2Al_2SiO_7$ side of the diagram than they are shown, as the preponderance of melilite in the charges made it impossible to determine the exact nature of the other phases present.

THE CRYSTALLIZATION OF A TYPICAL MIXTURE

Since the exact composition of the phases in equilibrium at any point in the diagram for this system is not known, the course of crystallization for a given mixture can be described only in a general way. Thus, in the case of a melt of composition 70 per cent $NaAlSiO_4$, 30 per cent $Ca_2Al_2SiO_7$, the first crystals begin to form at a temperature of 1323° C. These are carnegieite mix-crystals. They continue to separate out until the melt has cooled down to 1298° C., causing the liquid to become continually poorer in $NaAlSiO_4$ and in the material in solid solution with it. The carnegieite solid solutions presumably become more concentrated as the temperature falls. At the temperature indicated, the carnegieite begins to invert to nepheline. As the substance in solid solution in this phase is probably different from that entering into the carnegieite solid solutions, the latter is then slowly returned to the liquid solution while the former substance is progressively subtracted from it. Melilite begins to crystallize at 1278° C. and continues to do so in the presence of carnegieite, nepheline mix-crystals, and liquid until, at 1197° C., the inversion to nepheline is complete and the last car-

negieite disappears. At some lower temperature the last of the liquid will be used up.

THE SUBSTANCES IN SOLID SOLUTION

In the foregoing, frequent mention has been made of the fact that the crystalline phases of this system are solid solutions. The nature of the substances in solid solution is, unfortunately, not decided by the results of this study. However, in each case certain facts are known which suggest a choice among the several possibilities.

Nepheline.—In the case of nepheline, several substances are known to enter into solid solution with it in considerable amount. Of the molecules entering into the composition of natural nephelines, $KAlSiO_4$ is known to form a complete solid-solution series with nepheline,[12] while albite[13] and anorthite[14] are each taken into solid solution up to 35 per cent. No potash compound is to be considered a possibility in the present case, as the system contains no potash. Anorthite is the most likely of the other two possibilities, because it is known to change the refractive indices of nepheline considerably when in solid solution with it.

If anorthite is indeed the substance present in the nepheline solid solutions, certain aspects of its behavior are not quite clear. Bowen[15] found that nepheline solid solution with anorthite underwent a change in optic sign from negative to positive as the amount of the latter substance increased. The value of the index ω remained constant, while that of ϵ increased from 1.533 to 1.539. In the present study, a more marked rise in indices occurred, a value as high as 1.552 being reached. The birefringence decreased almost to zero, but it could not be determined whether a change of optical sign took place. Bowen[16] found that solid solution of anorthite in nepheline raised the temperature of the inversion to carnegieite as much as 100°, while in the present study the upper boundary of the inversion range is at most only about 50° above the normal inversion. In this case, however, solid solution in carnegieite is also considerable, and although the substance in solid solution is, as the general relations show, not anorthite, this may affect the extent to which the inversion is raised. Anorthite may be thought of as produced by an interaction between gehlenite and nepheline as expressed by the equation $Ca_2Al_2SiO_7 + 2NaAlSiO_4 \rightleftharpoons CaAl_2Si_2O_8 + Na_2Al_2O_4 + CaSiO_3$.

Carnegieite.—The carnegieite solid solutions present a similar problem. Clearly, in this case solid solution modifies the composition of carnegieite very considerably, as indicated by marked raising of the indices of refraction and by the lowering of the inversion boundary. Inversion lowering is a very unusual phenomenon in silicate systems. This is because the high-temperature forms of enantiotropic minerals usually form solid solutions much less readily than do the low-temperature modifications. Thus, the extent of solid solution of anorthite in carnegieite is only 5 per cent, while it is 35 per cent in nepheline.[17] However, two substances are known which form more concentrated solid solutions in carnegieite than in nepheline. The artificial mineral sodium metasilicate (Na_2SiO_3) was found by C. E. Tilley[18] to enter into carnegieite solid solutions up to 24 per cent

[12] N. L. Bowen, "The Sodium-Potassium Nephelites," *Amer. Jour. Sci.*, Vol. XLIII (4th ser., 1917), pp. 115–32.

[13] P. 106 of ftn. 5.

[14] P. 563 of ftn. 2.

[15] P. 566 of ftn. 2.

[16] P. 571 of ftn. 2.

[17] P. 563 of ftn. 2.

[18] The Ternary System, Na_2SiO_3-$Na_2Si_2O_5$-$NaAlSiO_4$, *Min. u. petrog. Mitt.*, Vol. XLIII (1933), p. 414.

but not at all into nepheline. The solid solution in carnegieite caused an increase in the birefringence of that phase by increasing the value of the index γ to 1.518. It also resulted in a wide inversion interval, which extended from the normal inversion down to 1163° C. In the present case, solid solution raises the indices of carnegieite much more noticeably and results in a marked decrease in birefringence. The inversion lowering is slightly greater than in the case cited above.

Sodium aluminate ($Na_2Al_2O_4$) has also been found to enter in considerable amounts into solid solution with carnegieite,[19] although the extent to which this can occur is not known as yet. Data regarding the optical and other properties of these solid solutions are also not yet available. Of the two substances considdered, sodium aluminate is perhaps the more likely constituent of the carnegieite solid solutions under consideration. As already noted, the trend of these solid solutions is to increase in concentration with increase in the $Ca_2Al_2SiO_7$ content of the mixture. As no suitable lime-bearing compound is known, the substance sought may be one whose formation is favored by the decrease in silica rather than by the increase in lime in this direction. Sodium aluminate is just such a compound. It may be formed in these melts by a reaction between nepheline and gehlenite such as the one cited above in which gehlenite and nepheline react to give $CaAl_2Si_2O_8$, $Na_2Al_2O_4$, and $CaSiO_3$.

It is hoped that future work will provide a basis for a more definite idea of the composition of these solid solutions.

Gehlenite.—In the case of the gehlenite solid solutions there is also little indication of what the substance modifying the composition actually is. In nature, gehlenite shows a strong tendency to form complex solid solutions, so that the compositions of the natural melilites require at least four hypothetical molecules for their expression by any one of the several theories which have been advanced to explain them. Assuming that only one substance besides gehlenite enters into the composition of these artificial melilites, and reasoning from the evidence of the optical properties, as was done in the case of carnegieite and nepheline, it may be considered probable that the substance in question is a soda compound. A compound analogous to gehlenite but containing some soda is $CaNaAlSi_2O_7$, one of the hypothetical melilite molecules suggested by Winchell.[20] This differs from gehlenite only in the substitution of Na for Ca, and Si for Al, replacements consistent with present knowledge of atomic volumes. This compound may be thought of as formed by a reaction between nepheline and wollastonite, as $NaAlSiO_4 + CaSiO_3 \rightleftharpoons CaNaAlSi_2O_7$. $CaSiO_3$ is one of the products of the reaction between nepheline and gehlenite cited above, by which $CaAl_2Si_2O_8$ (anorthite) and $Na_2Al_2O_4$ (sodium aluminate) are also formed. These reactions, while purely hypothetical, do serve in some measure to suggest in what ways the compositions of the liquids may change as the mix-crystals separate out.

PETROLOGIC SIGNIFICANCE

THE SODA-LIME-ALUMINA-SILICA SYSTEM

The present investigation deals with but a small part of a much larger system —the quaternary system soda-lime-alumina-silica Knowledge of the equilibrium relations in the parent-system is of great importance in the understanding of petrologic problems. This is evident

[19] Unpublished work of J. F. Schairer and N. L. Bowen.

[20] A. N. Winchell, "The Composition of Melilite," *Amer. Jour. Sci.*, Vol. VIII (5th ser., 1924), pp. 375–87.

from a consideration of the quantitative importance of these oxides in the compositions of igneous rocks. The four together constitute 83 per cent, by weight, of the average oxide composition of igneous rocks.[21] The petrologic significance of the system may be further illustrated by listing those minerals which consist of these four oxides alone. Among these are the plagioclase feldspars, quartz, nepheline, wollastonite, gehlenite, grossularite garnet, larnite, sillimanite, corundum, mullite, and jadeite. Mention may be made of the fact that the system of four oxides has importance also in a number of technological applications. Knowledge of equilibrium relationships in this system is of fundamental importance in glass technology, the manufacture of refractories, and Portland cement manufacture.

THE SYSTEM $NaAlSiO_4$-$Ca_2Al_2SiO_7$

Comparison with natural magmas.—The synthetic melts of this system do not coincide in composition with any natural magmas. Perhaps the closest approach is found in the magmas which formed certain feldspar-free rocks of the alkaline clan, such as the urtites and ijolites, nepheline- and melilite-basalts, and the turjaites and melilite-nephelinites. Of these, the latter two types, while less common, contain prominent amounts of both melilite and nepheline, while urtites have the highest proportion of nepheline —over 70 per cent. They all contain considerable amounts of alkaline pyroxenes and a variety of accessory minerals. Also, the melilites present all contain magnesia, and thus differ from those of the present system.

The significance of the solid solutions.—The results of this investigation serve to demonstrate once again the great tendency of both nepheline and gehlenite to form solid solutions, which makes it very unlikely that any mineral will ever be found in nature having the composition of either pure compound. The fact that pure nepheline cannot exist in the presence of a lime-bearing compound has been corroborated by the experiments of several investigators. The present study shows that pure gehlenite does not form if even a small proportion of a soda-bearing compound is present. In both cases, of course, this is just what would be expected from the complexity of the compositions of the natural minerals. However, the present system, with its simplicity of composition, might be expected to yield solid phases of simple composition where more complex melts would not. If the substance in solid solution with nepheline is indeed anorthite (or a plagioclase close to anorthite in composition), another illustration is provided of the great tendency of feldspar molecules to form even in melts of such low silica content. It suggests that plagioclase may well be present, in solid solution with nepheline, even in feldspar-free rocks such as the urtites and ijolites.

Geologic thermometry.—The value of the nepheline-carnegieite inversion as a "geologic thermometer" has been discussed by many earlier investigators. This study sheds some light on the question by furnishing an even lower value for the minimum temperature at which carnegieite can be stable. In this system, carnegieite was obtained in equilibrium at temperatures somewhat below 1150° C., or 100° lower than the inversion for the pure substance. The temperature range of the inversion was as great as 150°. It seems probable that a similar inversion interval could be expected in many natural nephelines, and, were they

[21] F. W. Clarke and H. S. Washington, "The Composition of the Earth's Crust," *U.S. Geol. Surv. Prof. Paper 127* (1924), p. 16.

crystallized from magmas at temperatures above 1150° C., some carnegieite might have formed. Since none has been found in nature, it is virtually certain that the magmas which formed the nepheline-rich rocks crystallized at temperatures lower than that.

In considerations of the use of inversions as geologic thermometers, the effects of pressure should be taken into account. In the present investigation, as is customary in such studies, experiments were carried out at atmospheric pressure. High pressures would significantly shift the nepheline-carnegieite inversion temperatures, because the densities of the two phases differ considerably. The tendency would be to raise the temperature of inversion, as nepheline is the more dense.

In most of the systems in which nepheline and carnegieite have been studied, nepheline will crystallize directly without the formation of any carnegieite in melts in which there is less than a certain concentration of $NaAlSiO_4$. That is, the compositions of the melts lie outside the field of carnegieite. In systems in which the nepheline takes large amounts of another component into solid solution, certain melts crystallize entirely to nepheline mix-crystals without the formation of any carnegieite. Certain natural nephelines are known which also have the property of melting and crystallizing directly.[22] The above facts lead to the conclusion that only in rocks very rich in $NaAlSiO_4$ could carnegieite be expected to form. However, the present study furnishes a case in which carnegieite is a stable phase on both sides of the "eutectic." It is formed in all melts containing more than 50 per cent $NaAlSiO_4$, and probably also in some poorer in that component. This suggests the possibility that some nepheline-rich magmas, if also rich in lime, might have similar wide "fields" of carnegieite, embracing compositions other than those especially high in $NaAlSiO_4$.

Unusual nature of the system.—Perhaps the chief significance of the system $NaAlSiO_4$-$Ca_2Al_2SiO_7$ lies in the illustration which it affords of the extent to which complexities can appear in a system which might be expected to be binary. Many investigated systems consisting of mixtures of two substances behave in true binary fashion, i.e., the compositions of all liquid and solid solutions can be represented by points lying along the binary join. In this system only three solid phases are present, two of them enantiotropic forms of the same mineral, but the composition of each diverges from the binary join in a different direction. As the composition of the phases can be represented only in three dimensions, the binary diagram gives little information about them. It does show some unusual relations between the phases, however. Already discussed are the indeterminate low temperature of complete crystallization, the wide field of stability of carnegieite, and the anomalous reversal of the normal relations between carnegieite and nepheline for certain compositions. The puzzling aspects of this system may be made clear as the result of work on adjoining compositions.

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[22] H. S. Washington and F. E. Wright, "A Feldspar from Linosa and the Existence of Soda-Anorthite (Carnegieite)," *Amer. Jour. Sci.*, Vol. XXIX (4th ser., 1910), p. 69.

AN EXAMPLE OF EXFOLIATION CAUSED BY CHEMICAL WEATHERING

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ABSTRACT

Farmin has contended that the primary cause of natural exfoliation of rock surfaces is dilation due to relief of pressure. This conclusion is questioned, and evidence from a mine adit in Sonora is presented which indicates that some exfoliation results primarily from chemical disintegration without the aid of pre-existing surfaces of strain due to dilation.

POSTULATED CAUSES OF EXFOLIATION

Of the many causes postulated to explain exfoliation of rocks in nature, perhaps most attention has been given to these four: (1) insolation, (2) forest fires, (3) chemical disintegration, and (4) dilation.

Insolation has been discarded by most geologists as a possible cause of any great importance because of failure of field or laboratory evidence to support it.[1,2,3]

That exfoliation may be caused by the heat generated by forest and brush fires has been demonstrated fairly conclusively,[4,5] but this one factor cannot explain all known examples of exfoliation.

Chemical disintegration in weathering has been considered by many geologists to be an important cause of exfoliation, but Rollin Farmin[6] has attempted to relegate it to the position of a secondary cause and to explain all exfoliation, save possibly that due to fires, as resulting primarily from dilation due to removal of load on rocks.

DILATION AS A PREREQUISITE OF EXFOLIATION

Some of the examples Farmin cites lend support to his hypothesis, and there can be little question that the sheet structure observed in many domes is due to relief from load.[7] On the other hand, the onionskin structure of the pebbles he describes in the Tintic pebble dikes could be due to contact with hot fluids rather than to release of pressure; indeed, it is difficult to visualize release of pressure as the primary agent in forming this structure in a small, hard quartzite pebble. Pebbles recovered from the bottoms of deep oil wells might furnish evidence on this point.

Donald Barton[8] has attempted to disprove Farmin's thesis by showing that exfoliation of granite in Egypt is better developed where the rock is subjected to

[1] Eliot Blackwelder, "Exfoliation as a Phase of Rock Weathering," *Jour. Geol.*, Vol. XXXIII (1925), pp. 793–806.

[2] Eliot Blackwelder, "The Insolation Hypothesis of Rock Weathering," *Amer. Jour. Sci.*, Vol. XXVI (1933), pp. 97–113.

[3] David T. Griggs, "The Factor of Fatigue in Rock Exfoliation," *Jour. Geol.*, Vol. XLIV (1936), pp. 783–96.

[4] Eliot Blackwelder, "Fire as an Agent in Rock Weathering," *Jour. Geol.*, Vol. XXXV (1927), pp. 134–40.

[5] K. O. Emery, "Brush Fires and Rock Exfoliation," *Amer. Jour. Sci.*, Vol. CCXLII (1944), pp. 506–8.

[6] "Hypogene Exfoliation in Rock Masses," *Jour. Geol.*, Vol. XLV (1937), pp. 625–35.

[7] Richard H. Jahns, "Sheet Structure in Granites," *Jour. Geol.*, Vol. LI (1943), pp. 71–98.

[8] "Discussion: The Disintegration and Exfoliation of Granite in Egypt," *Jour. Geol.*, Vol. XLVI (1938), pp. 109–11.

more moisture, but this argument does not disprove the existence of a primary structure due to dilation; the effect of moisture could still be only secondary. Where no exfoliation was noted, it would be possible to argue that invisible surfaces of strain did exist, awaiting the helping hand of hydration to develop into exfoliation.[9]

If it is granted that dilation may be one cause of exfoliation (as distinct from sheeting) of boulders and small outcrops, can it be considered the only primary cause in nature, aside possibly from forest fires? Omitting from discussion spheroidal structures in mine dumps—until an example is found of an exfoliated rock mass which can be proved never to have been under great load—one must rely on the general form and structure of certain exfoliated and spheroidal surfaces. It is common to observe granite outcrops with a hemisphere or small dome in each quadrilateral formed by the intersection of two joint sets[10] or rows of small domes between parallel joints. In many such outcrops the rounded forms are obviously the result of the modification of corners and edges produced by random joint intersections. Such general considerations lead one at once to question the wide applicability of Farmin's conclusion. Nevertheless, even for such examples, the proponent of dilation could put up arguments in favor of his theory which might be difficult to disprove positively, strain though they might the theory of probability. However, a mine adit examined by the writer in December, 1945, provides more convincing evidence on this question.

[9] Rollin Farmin, "Discussion: The Disintegration and Exfoliation of Granite in Egypt," *Jour. Geol.*, Vol. XLVI (1938), pp. 892–93.

[10] James Romanes, "Geology of a Part of Costa Rica," *Quart. Jour. Geol. Soc. London*, Vol. LXVIII (1912), p. 136.

SPHEROIDAL FORMS IN A SONORA ADIT

The Hollingsworth adit, driven in 1925 at an elevation of about 6,500 feet into the western face of Realito Mountain near the summit, lies about 9 miles southeast of Tarachi, Sonora, or 228 miles due south of Animas Station on the Southern Pacific Railway in New Mexico. The adit is straight, close to horizontal, and 223 feet long. It runs south 71° east from the portal and averages about 7 feet high and 5 feet wide. The rock exposed in the adit is the predominant rock of the region for miles around and is a porphyritic and pyroclastic volcanic of andesitic composition characterized by rather weak banded structure, mostly at dips less than 30°, and abundant rounded to subangular inclusions from an inch to a foot in diameter of rock of the same medium-grain size and appearance as the enclosing rock.

The greater part of the adit shows exfoliation to a marked degree; the walls. and back are smooth, rounded, and undulating, and the floor is covered to a depth of about 1 foot with the granular and scaly residue from the back and walls. The individual spheroidal knobs are from 1 foot to 3 feet in diameter, but the exfoliation has not proceeded to the point of producing forms which project more than about 8 inches out from the wall. Layers of crumbly scale, 0.02 inch to 0.10 inch thick and totaling several inches in thickness, cover much of the walls and to a less extent the back. These layers can be easily knocked off with a pick or even with the fingers, exposing very smooth, rounded, and fairly hard spheroidal bedrock underneath.

No exfoliation is found for the first 28 feet in from the mouth of the adit. From 45 feet to 65 feet the rounding is most highly developed, but with increased distance from the portal the individual

scales are found to be thicker, and beyond 65 feet many spalls 1 inch to 2 inches thick are present. Beyond 105 feet rounding is not well developed, and instead of scale there are spalls 0.25 inch to 5 inches thick. At 160 feet from the portal there is a strong seep of water from the back, and from this point to the face the walls and back are wet. No exfoliation and no rubble on the floor are found from 160 feet to 213 feet, and in this section the adit cuts rock showing considerably more hydrothermal alteration than the rest of the adit. From 213 feet to 223 feet exfoliation is again encountered.

In the length of adit where spheroidal weathering is best developed, from 45 feet to 65 feet, the rock is similar in every other respect to that of the rest of the adit up to 160 feet. The spheroids cut indiscriminately across included fragments and fracture planes, and the amount of alteration, which in all parts of the adit is distributed erratically, appears to have had no effect on the development of the rounded forms. Where the spheroidal weathering is less well developed, fractures and narrow mineralized seams affect the shape of the spheroids.

The western slopes of Realito Mountain, under which this adit was driven, experience less marked diurnal changes in temperature than the plateau country of Sonora at a similar elevation. In winter the range is from about 35° to about 65°; in summer it is about 60° to about 85°, with occasional extremes above and below these figures. The annual precipitation is 15 inches or less. As this adit has only one outside connection and enters a steep hill, most of it remains at a nearly constant temperature the year round, the extreme of variation probably being not more than one or two degrees Fahrenheit. The air temperature in most of the adit is probably about 70° F.; the relative humidity is probably between 70 and 90 per cent. After working several hours one night in December near the face of the adit, the writer walked out of the adit at 3:00 A.M. He first noticed a change in air temperature (a drop) at 28 feet from the portal, just at the limit of exfoliation going out.

THE HOLLINGSWORTH ADIT AND THE DILATION HYPOTHESIS

Insolation, of course, cannot be a factor in the exfoliation and spheroidal disintegration of these rock surfaces. Furthermore, there is no evidence that there has ever been a fire in the adit. Chemical disintegration and dilation are the only two processes which might have been operative to produce the forms seen, for there is nothing to suggest that the rounded forms have been inherited from a flow or pillow structure in the original rock; the rounded surfaces cut across included fragments and are not parallel to the banding of the rock even in the most general way.

That there has been chemical disintegration in the Hollingsworth adit is obvious. Has it followed, however, an earlier spheroidal structure caused by dilation? If the rounded surfaces in this adit were due to dilation, some of the knobs should approximate spheres or hemispheres, but they do not. Each knob is a small segment of a sphere, and immediately adjacent to it are other knobs, both smaller and larger. In other words, any knob which one may select and which one may in theory project into the rock wall as a sphere will intersect several other spheres thus projected. If, then, these curved surfaces are the reflection of an internal structure of the rock due to dilation, the dilation pattern of

the rock is unbelievably complicated, and it remains difficult to explain why some few of the partial spheres never developed into more nearly complete spheres. A simpler and more obvious explanation is that the surfaces are merely the modification of the irregular protuberances which were left after each blast on the walls and back of this adit as it was run. The influence of fractures and narrow seams on the shape of the knobs, particularly in those parts of the adit where the spheroidal weathering is less perfectly developed, supports this explanation. It is as if one were to run over the walls and back of an ordinary crosscut with sandpaper, smoothing out the bumps but not rubbing them entirely away. Would not this be the normal and expected result of the chemical action of moist air on an irregular surface of rock of the proper composition?

Farmin[11] mentions exfoliated structures of spheroidal type in a mine in Amador County, California, but I gather from his very brief description that what he observed were sections through spheroidal structures along directions the working happened to take, just as one may cut through a spheroidally exfoliated boulder in building a road, revealing concentric circles on the side of the cut. There is nothing of this description in the Hollingsworth adit. It is the conclusion of the writer, therefore, that the rounded surfaces seen in this adit in Sonora are due to the simple action of moist air on an uneven rock face.

SOME FACTORS INFLUENCING THE FORMATION OF SPHEROIDS

It might be supposed from the short time in which these surfaces have formed that the country rock in this region is particularly well adapted to the formation of spheroids, but spheroidal weathering is almost nonexistent in the many surface exposures of this same rock near by and for miles around. On the other hand, soft rocks as well as hard, and altered rocks as well as fresh, show good spheroidal forms in the adit.

There is probably, however, a limit to the differences in physical and chemical characteristics of this particular rock beyond which spheroidal weathering will not take place. The rock from 160 feet to 213 feet from the portal of the adit, which is well altered hydrothermally, does not show these forms. However, even here, this may be explained on another basis: the constant dampness of the walls and back of the adit in this section. These surfaces are probably damp the year round, for the writer visited the adit in a period when there had been no precipitation for many months. The rock in the rest of the adit is dry and is almost certainly in this condition the year round.

The conclusion seems inescapable that, given a rock which can form spheroidal surfaces upon weathering, such forms may be developed in a very few years under conditions of constant temperature merely by maintaining the proper humidity, and possibly temperature. Near the portal of the Hollingsworth adit the dry air from outside reduces the humidity of the air inside sufficiently to prevent the formation of spheroids in this particular rock. Near the face of the adit either the dampness of the surfaces or the type of rock alteration prevents the formation of the spheroids; the writer is inclined to consider the dampness to be the controlling factor, for parts of this highly altered zone contain rock that is quite hard and ap-

[11] Pp. 626–27 of ftn. 6.

parently similar to the rock elsewhere. In other words, moist air at a constant temperature appears capable of producing spheroidal weathering shapes on a dry rock exposure.

It must not be assumed, however, from this discussion that the right conditions of humidity and temperature alone can produce spheroidal disintegration in any rock. If this were the case, spheroidal forms would be extremely common in mines throughout the world, and such is far from the truth. What is needed apparently is a medium-grained to coarse rock which disintegrates into a sand. Rocks which slough off in chunks or in powder will not produce the spheroids; the chunks follow joints, the powder is not cohesive. In some granular rocks, however, given the proper conditions of disintegration, the concentric shells which tend to form are sufficiently cohesive to allow the rock to peel and to allow the alteration to progress uniformly into the rock along the most efficient geometrical front: a curved surface.

ELECTRODE POTENTIALS AND FREE-ENERGY CHANGES IN GEOLOGY

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ABSTRACT

By use of electrode potentials and free-energy changes the term "chemical affinity" may be placed on a quantitative basis. The significance and application of these two concepts to geology is briefly suggested. For example, one may explain more satisfactorily why certain elements occur most commonly in the free state, whereas others are found invariably in compounds.

INTRODUCTION

For many years it has been a common practice in geology to use the term "affinity" in reference to the mutual attraction of certain chemical elements. This term is strictly a qualitative one expressing the attraction between the element in consideration and other elements or radicals. It has been used to explain why certain elements invariably form compounds in nature whereas others occur commonly in the native state. For example: in explaining the nonoccurrence of sodium as a native element, it is said that sodium has such a great affinity for other elements that it cannot exist uncombined; on the other hand, it is concluded that platinum occurs as a native element by virtue of its low affinity for other elements and, therefore, its weak tendency to form compounds.

It is the purpose of this paper to show how the term "affinity" may be placed on a quantitative basis through the use of electrode potentials and free-energy changes. Since most geologists are unfamiliar with the details of the concepts of electrode potentials and free-energy changes, a brief, but adequate, treatment of these two concepts will be given here; and it will be pointed out how free-energy changes help to explain the chemical states in which we find the numerous chemical elements. Those interested in a more detailed treatment of these concepts are referred to the numerous publications listed at the end of this paper.

ELECTRODE POTENTIALS

If we were to construct a cell composed of a zinc electrode and a platinum electrode immersed in a solution of zinc chloride and hydrochloric acid and to connect the electrodes by an electrical conducting wire to which a voltmeter is attached in parallel, we could measure the electrical drop across the cell. As the electrons flow through the wire, the following reactions take place in the cell: (1) At the zinc electrode the metallic zinc becomes ionized, and electrons are set free. (2) At the platinum electrode ionized hydrogen is converted to hydrogen gas by the addition of electrons. These reactions may be written as follows:

(1) At the Zn electrode:

$$Zn = Zn^{++} + 2e^-$$

(2) At the Pt electrode:

$$2H^+ + 2e^- = H_2 .$$

Adding these two electrode reactions we get the total reaction taking place in the cell.

Total cell reaction:

$$Zn + 2H^+ = Zn^{++} + H_2 .$$

Any oxidation-reduction reaction may be split into two "couples" which indicate the manner in which the electrons are transferred. Using the example given the reaction

$$Zn + 2H^+ = Zn^{++} + H_2$$

may be written as two couples:

$$Zn = Zn^{++} + 2e^-$$

$$H_2 = 2H^+ + 2e^- .$$

Now, let us assign to the H_2(gas)–H^+(ion) couple an electrode potential (abbreviated E^0) value of zero. This is merely an arbitrary assignment, and we could just as well have assigned a zero electromotive force to the Zn–Zn^{++} couple. It is conventional, however, to assign the H_2–H^+ couple an electromotive force of zero. Hence we may say:

$$H_2 = 2H^+ + e2^- \qquad E^0 = 0.000 \text{ volts.}$$

If we now read the voltmeter attached to the cell described, we will obtain a value of

$$E^0 = 0.7620 \text{ volts.}$$

This electromotive force of 0.7620 volts is due to a combination of the Zn–Zn^{++} couple and the H_2–H^+ couple. Since the H_2–H^+ couple was assigned an electromotive force of zero volts, then it may be considered that the total electrical drop across the cell is due to the Zn–Zn^{++} couple. This may be stated:

$$Zn = Zn^{++} + 2e^- \qquad E^0 = 0.7620 \text{ volts.}$$

In a similar manner the potentials of other elements may be measured, but in many cases indirect methods must be used. Such determinations have been made for many of the elements, and the values given are for certain standard conditions. These conditions are: a temperature of 25° C., a fugacity (thermodynamic pressure) of one atmosphere for all gases involved, and an activity (thermodynamic concentration) of one molal (one mole per 1,000 grams of water) for all dissolved substances. Table 2 gives the values of E^0 for many of the metals. This list has been prepared from some of the more standard publications, and it is realized that the values given may not in all cases be from the most reliable source; however, they are sufficiently accurate for the present purpose. The electrode potentials have not been determined for many elements notable among which are most of the rare earths and, of course, the newly discovered transuranic elements.

SIGNIFICANCE OF ELECTRODE POTENTIALS

It is conventional to write all couples with the electrons on the right-hand side of the equation. Thus a positive value of a potential for a couple like

$$Ca = Ca^{++} + 2e^- \qquad E^0 = 2.87 \text{ volts}$$

means that the reaction

$$Ca + 2H^+ \rightarrow Ca^{++} + H_2$$

will proceed from left to right, as written. On the other hand, a negative value of E^0 implies that the reaction will proceed from right to left. For example:

$$Ag = Ag^+ + e^- \qquad E^0 = -1.42 \text{ volts}$$

means that the reaction

$$Ag + H^+ \leftarrow Ag^+ + \tfrac{1}{2}H_2$$

goes from right to left, as written.

These reactions may now be stated more simply as follows:

1. A reaction with a positive electrode potential tends to proceed from left to right.
2. A reaction with a negative electrode potential tends to proceed from right to left.

FREE-ENERGY CHANGES

In order to compare quantitatively the affinities of elements we must introduce the free-energy concept. The total energy content of a body is considered to consist of bound and free energy. Bound energy may be involved in such changes as those of state, whereas the free energy alone may be capable of transformation into other forms of work. It can be proved theoretically, however, that there must be a decrease of free energy in any reaction proceeding of its own accord.

In a reaction the bound energy may either decrease (be partly converted into heat), remain unchanged, or increase. If the decrease of free energy in a reaction is greater than the increase of bound energy, the reaction is exorthermic; but if the decrease of free energy is less than the increase of bound energy, the reaction is endothermic. Therefore, it is important to know the free energy gained or lost in a reaction rather than whether the effect is exothermic or endothermic. It might be added, however, that frequently the difference between the free energy and the total energy coming into play in a reaction is not great.

In conclusion, it is to be noted that substances tend to exist in the state which has the lowest free energy. Hence, in a reaction which involves a decrease in free energy the reaction tends to go to the right. For example, in the reaction

$$Na = Na^+ + e^-$$

the sodium ion has a lower free energy than the sodium atom, and so the reaction goes toward the right. Conversely, a reaction involving an increase in free energy will proceed from right to left, as this direction favors the formation of the state with the lower free energy.

The relation between electrode potentials and free-energy changes may now be stated:

$$\Delta F^0 = -n\mathfrak{F}E^0 ,$$

in which ΔF^0 is the change in free energy involved between the two sides of a couple, n is the number of electrons involved in the couple, $\mathfrak{F}$ is a constant having a value of 23,070 calories, and E^0 is the electrode potential of the couple. From this formula it is obvious that a negative electrode potential will give a positive free-energy change, both favoring a reaction from right to left; conversely, a positive electrode potential will give a negative free-energy change, favoring the reaction from left to right. This relationship is summarized in Table 1. It is to be borne in mind that electrode potentials should not be quantitatively compared, but changes in free energy must be used in making actual quantitative comparisons.

TABLE 1

REACTION DIRECTIONS

Sign of E^0	Sign of ΔF^0	Direction of Reaction
+	−	→
−	+	←

GEOLOGICAL APPLICATIONS

The low affinity possessed by many metals which occur in the native state and the great affinity of metals which are found in a combined state can be expressed quantitatively and understood more easily if we refer to Table 2. By

our previous reasoning we would not expect those metals exhibiting positive electrode potentials, and hence negative free-energy changes, to occur free in nature because they tend to form ions (i.e., the reactions go from left to right), causing them to enter into compound formation. It is obvious that these metals generally do not occur free. In a like manner, with all metals showing a negative electrode potential and hence a positive free-energy change, the tendency of the reaction is from right to left, which is against compound formation. This leads us to expect these substances to exist in the native state, which is generally found to be the case.

TABLE 2

ELECTRODE POTENTIALS AND FREE-ENERGY CHANGES

Couple	E° (In Volts)	ΔF° (In Calories)
$Ba = Ba^{++} + 2e^-$	2.90	−133,800
$Sr = Sr^{++} + 2e^-$	2.89	−133,300
$Ca = Ca^{++} + 2e^-$	2.87	−132,400
$Al = Al^{+++} + 3e^-$	1.67	−115,600
$Mg = Mg^{++} + 2e^-$	2.34	−108,000
$Ti = Ti^{++} + 2e^-$	1.75	− 80,750
$Li = Li^{+} + e^-$	3.02	− 69,670
$Cs = Cs^{+} + e^-$	3.02	− 69,670
$Rb = Rb^{+} + e^-$	2.99	− 68,980
$K = K^{+} + e^-$	2.92	− 67,400
$Na = Na^{+} + e^-$	2.77	− 63,590
$Cr = Cr^{+++} + 3e^-$	0.71	− 61,860
$Mn = Mn^{++} + 2e^-$	1.05	− 48,450
$Zn = Zn^{++} + 2e^-$	0.76	− 35,100
$Fe = Fe^{++} + 2e^-$	0.44	− 20,300
$Cd = Cd^{++} + 2e^-$	0.40	− 18,460
$Co = Co^{++} + 2e^-$	0.28	− 12,840
$Ni = Ni^{++} + 2e^-$	0.25	− 11,540
$Sn = Sn^{++} + 2e^-$	0.14	− 6,500
$Pb = Pb^{++} + 2e^-$	0.13	− 6,000
$Fe = Fe^{+++} + 3e^-$	0.036	− 2,530
$H_2 = 2H^{+} + 2e^-$	0.000	0
$Cu = Cu^{+} + e^-$	−0.552	12,040
$Cu = Cu^{++} + 2e^-$	−0.348	15,900
$Ag = Ag^{+} + e^-$	−0.780	18,450
$2Hg = Hg_2^{++} + 2e^-$	−0.799	36,900
$Pd = Pd^{++} + 2e^-$	−0.83	38,300
$Au = Au^{+} + e^-$	−1.68	38,700
$Hg = Hg^{++} + 2e^-$	−0.85	39,200
$Pt = Pt^{++} + 2e^-$	−1.2	55,000
$Au = Au^{+++} + 3e^-$	−1.42	98,000

Two logical conclusions may now be stated:

1. The more negative the change in free energy (calculated from $\Delta F^\circ = -n\mathfrak{F}E^\circ$) of an element, the greater the tendency toward compound formation and, theoretically, the less the probability of finding it free in nature.

2. The more positive the change in free energy (calculated from $\Delta F^\circ = -n\mathfrak{F}E^\circ$), the less the tendency toward compound formation and, theoretically, the greater the probability of finding it free in nature.

The writers do not wish to imply that the position of a metallic element in Table 2 is an infallible indication of its affinity for other elements. The elements are arranged in order of free-energy changes, which in turn are related to electrode potentials, as already explained. The conditions for which these values apply are: a temperature of 25° C., a fugacity of one atmosphere for all gases involved, and an activity of one molal for all dissolved substances. Under different conditions of temperature, confining pressure, and concentration the values of E° and ΔF° will differ; and the relative positions of certain elements may change. For purposes of comparison, however, data applicable to standard conditions seem logical and proper.

Although the table does not contain values for anions, the effects of these elements in compound formation cannot be ignored. Even though a particular cation may have a strong positive value for ΔF°, there may be an anion, with a strong negative value for ΔF°, available in nature and ready to combine with it. The writers, however, have intentionally omitted the anions from the table for the sake of simplicity. It is their intention to consider the effects of the anion as well as of the cation in compound formation in a succeeding paper.

Again, the values in Table 2 apply to

water solutions, and obviously should apply only to water-formed minerals.

At best, then, the table is actually a guide, because all reactions must be calculated or determined independently and cannot always be predicted. After all, the values for free-energy change are indicative of conditions at equilibrium; they show whether or not a given reaction may take place. They indicate neither the rate of reaction nor the length of time necessary to attain equilibrium.

In conclusion, it appears that the affinity of metallic elements may be expressed more quantitatively on the basis of free-energy changes. Table 2 indicates, therefore, why elements like gold, platinum, and palladium almost invariably occur free in rocks of the earth's crust; why silver, copper, and mercury less commonly occur free; why iron, lead, tin, nickel, cobalt, and zinc are rarely found in the native state; and why elements near the top of the list invariably form compounds. It must be emphasized, however, that there are numerous cases in which Table 2 apparently does not apply. Many of these will be considered in detail in a forthcoming paper.

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A TECTONIC INTRUSION OF SHALE IN ROCKBRIDGE COUNTY, VIRGINIA[1]

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ABSTRACT

A strange pipe-like mass of black shale and limestone in dolomite is exposed on U.S. Route 11 about 4 miles south of Lexington, Virginia. This mass is probably a tectonic intrusion caused by intense folding.

A peculiar mass of contorted black shale and gray crystalline limestone included in dolomite is exposed in a cut on U.S. Route 11 about 4 miles south of Lexington, Rockbridge County, Virginia. On the east side of the highway the mass is about 15 feet wide at the base and 7 feet wide at the top of the cut, which is about 15 feet deep (Fig. 1). Across the highway the shale-limestone mass is about 70 feet wide through the whole exposure. Shale, on the east side of the highway, is in direct contact with dolomite except at road level, where a small wedge of limestone intervenes on the north side. On the west side of the highway, however, 55 feet of crystalline limestone separates the shale from the dolomite to the north but is absent to the south (Pl. I). Continuous outcrops in a stream that flows parallel to, and about 50 feet west of, the road include no shale and no crystalline limestone in strike with the shale-limestone mass in the road cut. Furthermore, east of the highway there is no evidence of the shale-limestone mass in the bedrock. It is thus reasonably certain that the mass is a lenticular body about 100 feet long and not more than about 70 feet wide (Fig. 1).

Within the shale-limestone mass the rocks are so complexly deformed that, except along the contact of the shale and limestone, stratification cannot be traced continuously through the mass. Around the mass, however, the strata seem to be in normal sequence, though locally overturned. They include, from oldest to youngest, the Beekmantown, Mosheim(?) Holston(?), and Athens formations.[2] A disconformity with a relief of more than 100 feet occurs in places above the Beekmantown formation.[3] The Ordovician in the Lexington area consists, from the base upward, of the Chepultepec, Beekmantown, Murfreesboro, Mosheim, Lenoir, Holston, Whitesburg, Athens, Chambersburg, and Martinsburg formations.[4] No rock of typical Murfreesboro or Lenoir lithology has been positively identified in the vicinity of the shale-limestone mass south of Lexington. Limestone similar to the Mosheim occurs in small discontinuous masses along the boundary between the Holston-type limestone and Beekmantown formation, but in most places the Holston-type limestone lies directly upon the Beekmantown. Furthermore, the thickness of this

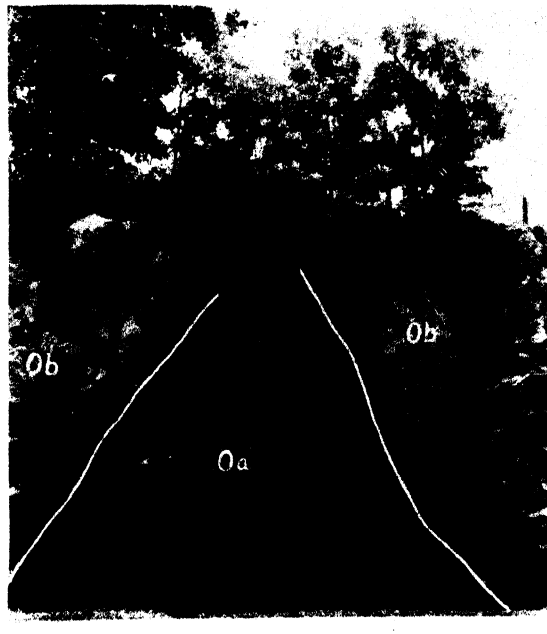
[1] Published with the permission of the state geologist of Virginia.

[2] Charles Butts, "Geology of the Appalachian Valley in Virginia," *Va. Geol. Surv. Bull. 52* (1940), Part I, pp. 102–19, 135–39, 148–54, 159–70.

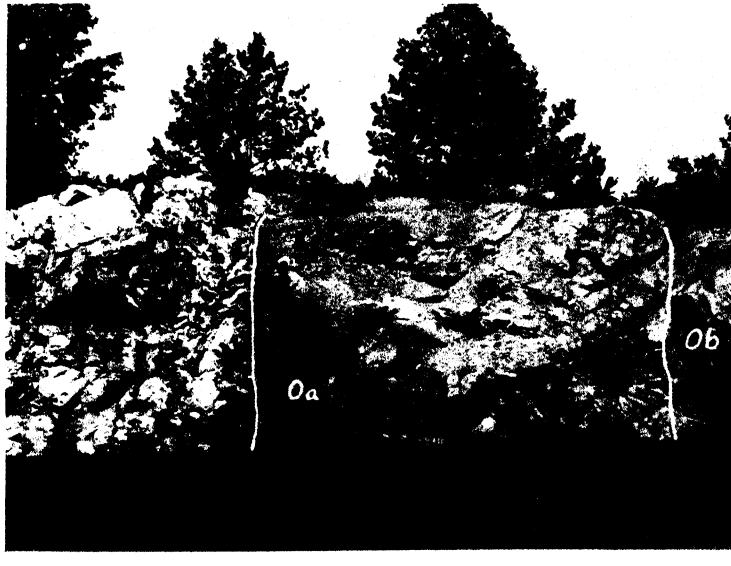
[3] *Ibid.*, p. 119. See also Byron N. Cooper, "Geology and Mineral Resources of the Burkes Garden Quadrangle, Virginia," *Va. Geol. Surv. Bull. 60* (1944), pp. 31–34.

[4] Charles Butts, "Geologic Map of the Appalachian Valley of Virginia with Explanatory Text," *Va. Geol. Surv. Bull. 42* (1933), map in pocket.

PLATE I



A



A, Tectonic intrusion of Athens shale (*Oa*) in Beekmantown dolomite (*Ob*) on the east side of United States Route 11, about 4 miles south of Lexington, Virginia. *B*, Tectonic intrusion of Athens shale (*Oa*) and Holston limestone (*Oc*) in Beekmantown dolomite (*Ob*) on the west side of United States Route 11, about 4 miles south of Lexington, Virginia.

PLATE II

The overturned contact of the Athens shale (*Oa*) and Holston limestone (*Oc*) on United States Route 11 about 300 feet north of the tectonic intrusion.

limestone varies from a minimum of 13 feet to a maximum of more than 100 feet. Just east of the shale-limestone mass the contact of the Beekmantown and younger limestone is decidedly irregular, with pinnacles or tongues of the older formation projecting a hundred feet or more into the limestone. The character of the post-Beekmantown erosion surface clearly indicates solution or karst erosion, at least on a small scale, prior to the deposition of the overlying strata. The absence of the Murfreesboro, the discontinuity of the Mosheim-like masses, the variable thickness of the Holston-type limestone, and the tongue-like projections of the Beekmantown into younger formations are evidence of the disconformity at the top of the Beekmantown formation. The absence of the Lenoir formation, normally above the Mosheim formation, cannot be explained by this unconformity. The so-called Mosheim in the vicinity of the shale-limestone mass has been identified on the basis of lithology only. Consequently, this vaughnitic rock, in places between the Holston-type limestone and the Beekmantown, may not be Mosheim and may actually represent a unit younger than the Lenoir. If this interpretation

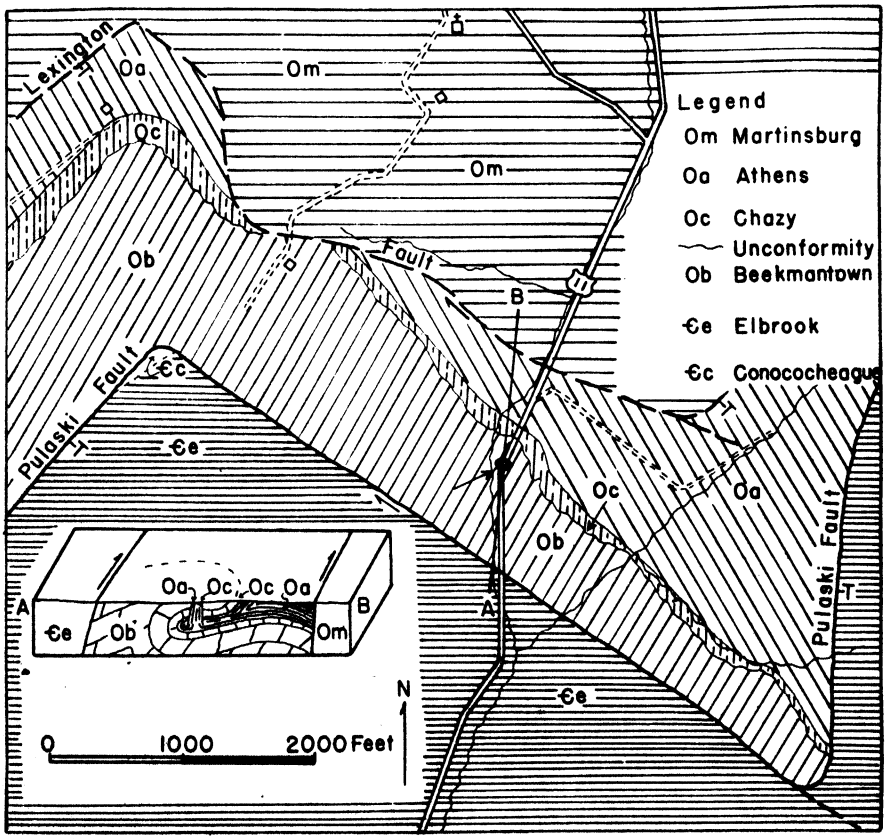


FIG. 1.—Geologic map of a small area 4 miles south of Lexington, Virginia

is correct, the unconformity above the Beekmantown may represent a hiatus that includes the Murfreesboro, Mosheim, and Lenoir formations. In other words, the Stones River group may be entirely missing at this locality.

The Holston-type limestone in the vicinity of the shale-limestone mass is stratigraphically overlain by black shale which is probably part of the Athens formation. The contact of these formations, where exposed on U.S. Route 11 about 300 feet north of the shale-limestone mass in the Beekmantown dolomite, is overturned. The black shale in the shale-limestone block included in the section of Beekmantown dolomite in the road cut is almost certainly the same formation as the shale 300 feet to the north. The limestone in the shale-limestone block is probably Holston limestone.

The relations of the small shale-limestone mass to the surrounding dolomite suggest faulting. B. N. Cooper and W. R. Brown,[5] of the Virginia Geological Survey, suggested that the shale-limestone block was dropped along gravity faults in an anticline. The major structure in which the block is contained is an anticline with the Elbrook formation thrust faulted upon the south limb, Beekmantown dolomite in the core, and the Holston and Athens formations on the north limb. The north limb of this anticline is overturned 300 feet north of the shale-limestone block (Pl. II). If the shale-limestone block were dropped along peripheral gravity faults, it is difficult to account for the relations of the limestone and shale in the block without postulating several faults. Furthermore, the severe cataclastic deformation of the rocks in the block indicates strong squeezing which seems incompatible with a structure produced by tension and gravity. If the block moved downward along gravity faults, the mass would probably have undergone little compression.

The attitude of the strata, the local stratigraphy, and the structure within the block suggest that the block moved into the Beekmantown dolomite from below. Just 300 feet northward across the strike from the shale-limestone block, the strata are overturned. Thus the Holston and Athens on the north limb of this overturned structure may be folded under the older Beekmantown dolomite. It seems possible that the Holston limestone and Athens shale might have been squeezed in the trough of an overturned syncline and forced upward along a shear zone into the older Beekmantown formation. The comparatively incompetent shale and limestone may have "flowed" upward and thus formed a tectonic intrusion into the Beekmantown. The boundaries of the block emplaced in the Beekmantown dolomite are thus local faults that extend downward only into the trough of an underlying overturned syncline (Fig. 1).

The anticlinal structure that contains the tectonic intrusion strikes about N. 35° W. This structure is flanked by a well-established fault on the south and a postulated fault on the north. These faults strike about N. 50° W. and offset the northeast-trending traces of the Pulaski overthrust fault and a postulated overthrust fault tentatively named the Lexington fault. The offsets in the traces of these faults are about a mile long. They may be caused by folding of the major fault planes or by tear faults. They cannot be attributed to the erosion of low-angle faults that dip uniformly toward the southeast. If the offsets are due to folding of the fault planes, these folds must be anticlines that strike about N.

[5] Personal communication.

20° W. The dip of the limbs of these folds should be of about the same order as the dip of the major faults along the northeast-trending traces because the limbs of the folds are highly divergent. However, the fault planes marked by the offsets dip at high angles, whereas the fault planes marked by the northeast-trending traces are low-angle. This marked difference in the magnitude of the dips and the orientation of the postulated folds across the regional strike (about N. 35° E.) casts some doubt upon the occurrence of fault-plane folds. The offsets may consequently be attributed to tear faults. It is difficult to prove the occurrence of such tear faults because the mapping units in the respective fault blocks are very thick and have broad outcrop belts.

If the faults that flank the northwest-trending structure containing the tectonic intrusion are tear faults, the relations may not be coincidental. Possibly the northwest-trending strata owe their orientation to the same differential stresses that caused the tear faults to develop. In other words, the northwest-trending strata may have been drag-folded against the tear in the Lexington fault block. Such drag-folding would cause the thinning of the units in the northwest-trending belt. The Holston formation is notably thinned in the hypothetical drag-folded zone.

PRIMARY LINEATION IN FLUVIAL SANDSTONES A CRITERION OF CURRENT DIRECTION[1]

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ABSTRACT

This paper describes primary current lineation of fluvial sandstones, a type of structure thought to be produced through the streaming of sand grains under the influence of shallow, smoothly flowing water. Field occurrence, validity, and use of the feature are briefly discussed.

Information as to the direction of movement of sediment-transporting currents of air or water is highly desirable in many geologic studies, especially those involving paleogeography. Direct observation, supported by theoretical considerations, has established several criteria which are listed by Frederic H. Lahee[2] as follows: cross-bedding; ripple mark; wave mark; rill mark; and, to some extent, the attitude of included fossils. All of these are megascopic features observable directly in the field. E. C. Dapples and J. F. Rominger[3] have made laboratory and statistical studies which indicate that current direction may be deduced from statistical analysis of measurements of the projections of grains on the bedding plane. They conclude that grains from laboratory fluvial and eolian environments exhibit a pronounced preferred elongation parallel to the direction of flow of the depositing medium and a marked tendency to lie with their larger ends upcurrent. Earl Ingerson[4] has shown in connection with a study of ripple marks that there is a possibility that petrofabric analysis may also be useful in some instances. The two last-mentioned methods require laboratory study with oriented specimens and presumably are most useful where all the usual megascopic evidence is lacking.

The purpose of the present paper is to describe another common feature of fluvial sandstones, here termed "primary current lineation," which may be used in the field as a supplemental criterion of current direction. The term "primary lineation" is used to distinguish this type of structure which originates at the time of deposition of the rock from possible linear structures which may originate through tectonic movements and which are more properly designated as secondary. It consists essentially of a streamlining or streaming effect of sand particles in relatively low and poorly defined windrow-like ridges parallel with the current direction. It is made visible mainly through the light and shadow effects of slight variations in relief upon the bedding planes. Figures 1 and 2 are intended to show the nature of the markings and to illustrate the effect of different angles of illumination in rendering them more noticeable. It should not be inferred that low-angle illumination is essential to the

[1] Published with the permission of the director, Geological Survey, United States Department of the Interior.

[2] *Field Geology* (4th ed.; New York and London: McGraw-Hill Book Co., 1941), p. 98.

[3] "Orientation Analysis of Fine-grained Clastic Sediments: A Report of Progress," *Jour. Geol.*, Vol. LIII (1945), pp. 246–61.

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FIG. 1.—Specimen of fluvial sandstone from Salt Wash sandstone member of Morrison formation, Carrizo Mountains, Arizona, showing current lineation. (Specimen photographed under strong oblique illumination coming from above and making an angle of about 10° with the surface. Black strip indicates inferred line of current direction.)

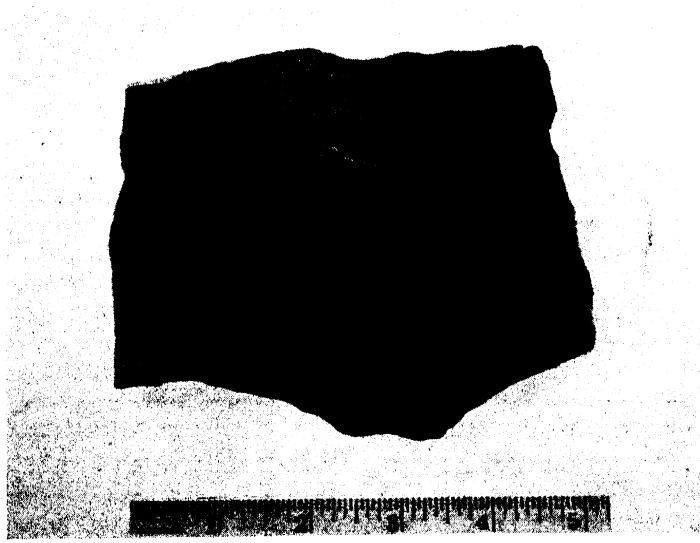

FIG. 2.—Same specimen as Fig. 1, photographed under vertical illumination. Note that lineation is still visible.

use of this feature; lineation is much more evident in gross aspect on larger exposures of sandstone than on small specimens.

Primary lineation in fluvial sandstones is thought to originate in shallow water of moderate velocity and, therefore, is characteristic of laminar flow and absence of turbulence except on a minute scale directly associated with the movement of individual sand grains. Sediments showing this feature are usually divided by flat bedding planes into laminae ranging from $\frac{1}{4}$ to $\frac{3}{4}$ inch in thickness. The same markings and lamination of beds usually persist through several inches or several feet of beds. Cross-bedding and current lineation occur together in sandstone lenses, but lineation is mostly confined to the marginal areas. Cross-bedded laminations are curved and inclined; those showing lineation are flat.

The validity of this criterion has been tested in the field by a variety of observations. Current lineation has been seen on the dry, sandy beds of intermittent streams in the arid regions of western Colorado, but it is apparently rarely formed and is distinctly subordinate to ripple markings in environments of this sort, probably because the flow of such streams is too swift and too turbulent. In all cases it is parallel with the direction of flow of the streams and at right angles to the ripple marks. In the fluvial sandstones of the Salt Wash sandstone member of the Morrison formation, current lineation has been checked many times and found to be in perfect agreement with the evidence of current bedding. It has been noted in fluvial sandstones of several different terrestrial formations of Paleozoic and Mesozoic age at many places in the Colorado Plateau and is assumed to be geologically widespread.

Current lineation, at least megascopically, limits the depositing current to two possible opposing directions and in this respect is analogous to flow structures in igneous rocks. Which direction is the correct one must be ascertained by reference to other evidence, usually current bedding. This is not a serious deterrent to the use of current lineation, since any conclusions regarding source, direction of flow, and general nature of streams responsible for fluvial deposits should be made on the basis of statistical analysis involving numerous observations. Isolated and scattered readings are likely to be entirely misleading and erroneous. For this reason all possible field evidence should be recognized and used. It has been the experience of the writer in making a detailed study of the sedimentary features of the Salt Wash sandstone member of the Morrison formation that the use of current lineation increases the potential number of observations which may be made in any one area by at least 30 per cent. In many small outcrops it is the only visible clue; a few inches of exposed rock are usually all that are needed for a good lineation reading, while several square feet or even yards must be visible for current-bedding determinations. Although current bedding is undoubtedly more useful and easier to interpret, current lineation is a valuable adjunct in studies of this kind.

REVIEWS

Our Oil Resources. Edited by LEONARD M. FANNING. New York and London: McGraw-Hill Book Co., Inc., 1945. Pp. 331. $4.00.

Articles and contributors: "The Role of Private Enterprise in the Development of Oil Resources," John A. Brown, late president, Socony-Vacuum Oil Company, Inc.; "American Oil Companies in Foreign Petroleum Operations," Eugene Holman, president, Standard Oil Company (New Jersey); "Conservation of Our Oil and Gas Resources," J. C. Hunter, president, Mid-Continent Oil and Gas Association; "Exploration Technology," O. D. Donnell, president, Ohio Oil Company, and A. Jacobsen, president, Amerada Petroleum Company; "Production Technology," John M. Lovejoy, president, Seaboard Oil Company; "Refining Technology," Robert E. Wilson, chairman of the Board, Standard Oil Company (Indiana); "The Earth's Petroleum Resources," Wallace E. Pratt, vice-president, Standard Oil Company (New Jersey); "Estimate of United States Oil Reserves," American Petroleum Institute; "Our Natural Gas Resources," Lyon F. Terry, second vice-president, Chase National Bank, New York; "Cost of Storage of Military Reserves of Crude Petroleum or Products," F. W. Abrams, vice-president, Standard Oil Company (New Jersey); "Our Reserves of Coal and Shale," K. C. Heald, chief geologist, Gulf Oil Corporation, and Eugene Ayres, chief chemist, Gulf Research and Development Company; "Oil from Coal and Shale," Robert E. Wilson, chairman of the Board, Standard Oil Company (Indiana); "The Public Domain and Naval Reserves," W. H. Ferguson, vice-president, Continental Oil Company; "Lands of the United States," A. C. Mattei, president, Honolulu Oil Corporation; "Status of Federal Lands," L. T. Barrow, vice-president, Humble Oil and Refining Company; "Capital Employed in the Petroleum Industry," Joseph E. Pogue, vice-president, and Frederick C. Coqueron, petroleum analyst, Chase National Bank, New York; and "The American Oil Industry," Leonard M. Fanning.

The problem of the amount of our oil resources is not simple. That it cannot be stated in terms of barrels or of years' supply is well brought out by this symposium written by as authoritative a group of authors as it would be possible to assemble. Several of the articles have been previously published, and there is some repetition of ideas and material, as might be expected in a series of articles such as this. The over-all result, however, is a fair picture of the complex nature of our oil and gas resources by a group of executives actively and successfully engaged in the discovery, production, and refining of oil and gas. The volume is well worth reading by anyone interested in the oil industry and the nature of its resources.

The general theme of the volume is optimistic as to the resources of oil and gas available for our future needs, *provided* that the oil industry is permitted to continue operating under an economy of free enterprise. These resources within the United States consist of the known reserves on the order of twenty billion barrels of crude petroleum and 110 trillion cubic feet of natural gas; the undiscovered oil and gas reserves; improved production methods, refinery techniques, and more efficient use of the products; the coal which can be transformed into gasoline and other products, chiefly by the Fischer-Tropsch process; and the oil shale which will yield large quantities of oil products through destructive distillation. The chief resources outside the United States, which may become available through import, are the large discovered and undiscovered petroleum reserves in many parts of the world, chiefly the Near East, the East Indies, and northern South America, a substantial percentage of which are now owned by American companies. Above all, however, "our greatest, though intangible, oil resource lies in the ability, energy, and brains of operating, engineering, and research talent working coordinately in the American private oil industry."

Numerous tables and charts show the essential data in convenient form. Of particular interest, perhaps, are those charts showing the steady expansion of the industry in the United States through good times and bad and whether we were at war or peace. Fundamentally, the expansion was based on the ability of the industry to discover enough oil to meet the demand,

and this ability to discover is dependent on many interrelated techniques and skills. Basically, discovery is geological in approach, but the methods by which the information is obtained—such as improved logging of wells, the use of geophysical methods for structural mapping, the advent of better steels, thereby making deeper drilling possible—all contributed to the result. These techniques and skills were developed largely because of the favorable political and economic climate in which the industry has operated, and one of the essentials of the continued expansion of the industry is in maintaining enough stimulation of effort to justify the drilling of the thousands of exploratory wells needed every year.

Not only have our oil resources increased through discovery of new reserves over the years but they have been multiplied even more by improved technologies and engineering skills in their production, manufacture, and use. Thus, improved production methods since 1918 are estimated to have added 35 per cent, or seventeen billion barrels, to the oil recovery formerly expected from the amount of oil discovered to date. Of the present estimated known reserves of twenty billion barrels, approximately eleven billion barrels, or 55 per cent, are estimated to be available through improved technology in drilling and producing alone. In a like manner, improved techniques in refining have increased the gasoline yields from around 25 per cent in 1918 to around 45 per cent in 1941, and it is anticipated that when the modern catalytic cracking-units have been generally installed, the gasoline yields from crudes will be near 57 per cent and limited only by the public demands for other products. The fear is expressed that the danger of the future will not be a shortage of raw materials but rather the devitalization of the entire research program through increased governmental domination, regimentation, and interference with the free play of technology and competitive enterprise.

The American oilman has found oil and gas in adequate amounts within the United States, and he has also played a leading and vital part in the exploration and development of oil and gas resources in many foreign lands. As a result of his aggressive policy, in 1939–41 American companies owned (excepting in the U.S.S.R.) approximately 23 per cent of the foreign refineries, 23 per cent of foreign oil production, and over 24 per cent of foreign oil reserves. These foreign reserves have been increased, until in 1944 it was estimated that the amount of American ownership is nearly 40 per cent, or a total of over ten billion barrels. This is nearly one-half the reserves known in the United States.

The record of gas resources runs parallel to that of oil. They are estimated by the Petroleum Administrator for War to be on the order of 110 trillion cubic feet of known producible reserves, or twenty-seven to thirty-three times the rate of annual withdrawal. This is considerably less than the later figure of 140 trillion cubic feet given by De Golyer before a committee of the Senate, June 19, 1945. As with oil, it is believed that the greatest future gas resource lies in the undiscovered reserves. The basis for this belief is in the geological interpretation of the unexplored regions of the country, which for both oil and gas are expected ultimately to develop at least as much more as has been found in the past.

One continuing problem of the oil industry has been conservation. Much progress has been made in this direction through the creation of the Interstate Compact in 1935. It now has a membership of thirteen states and includes more than 80 per cent of the annual oil production. The essence of modern conservation practice is the realization that an oil and gas reservoir is a physical unit; and, when its energy is controlled and utilized as a whole, the result is the recovery of a greater amount of oil from a given pool. Oil, having no inherent energy to lift it to the surface, is dependent on either (1) the expansion of gas dissolved in it, (2) displacement of oil by the influx of water from below, or (3) displacement of the oil by downward expansion of the free gas cap on the pool. In wells which have reached the pumping stage, gravity brings oil from the reservoir rock into the borehole. Thus conservation—"the prevention of the waste of an urgently needed natural resource"—is becoming more and more solidly based on sound engineering concepts of the conditions which prevail in the oil and gas reservoir.

While satisfactory progress is made in the field of conservation, the symposium expresses alarm over the effect of the increasing amounts of land being acquired by the federal government and thereby being withdrawn from oil and gas exploration. In amount it ranges to over 45 per cent of the lands in some of the western states and from 1 to 8 per cent for most of the oil-producing states east of the Rockies.

Several of the articles are of particular interest to geologists. That by John M. Lovejoy is a concise statement of the fundamental principles involved in the production of oil and gas and the relation of these phenomena to the geological conditions surrounding the reservoir. These principles, chiefly developed by petroleum engineers, have permeated the thinking of the entire industry on production and conservation problems.

Wallace Pratt, speaking from the vantage point of an executive of the largest oil company in the world, again reiterates his philosophy of the abundance of undiscovered oil and gas throughout the world and concludes that "there will be no sudden failure of our supply of petroleum. . . . . We shall have timely warning. The new industry of making liquid fuels from other raw materials will establish itself as rapidly as the petroleum supply fails . . . . it is hardly conceivable (unless wars are to continue unceasingly) that the United States will suffer a critical shortage of liquid fuels within the foreseeable future."

The overshadowing importance of coal as a possible substitute oil resource is brought out in the article by K. C. Heald and Eugene Ayres. The coal resources of North America are such that when gasoline and other products obtained from them compete in price with petroleum products, domestic requirements will be met for generations. The most likely method of obtaining gasoline and other products from coal is in the Fischer-Tropsch process, which has been extensively used in Europe but which so far has not been commercially utilized in the United States. This process uses two steps: (1) the gasification of any kind of coal or lignite and (2) the catalytic conversion of the synthesized gas to petroleum products. It may also be used directly on natural gas, eliminating step 1.

Since any kind of coal may be used in the Fischer-Tropsch process, there are on the order of four trillion tons of known coal source material in North America from which it is estimated some six trillion barrels of gasoline might be manufactured. This compares with an estimated fifty billion barrels of gasoline from oil shale, two billion barrels of gasoline from tar sands, two to four billion barrels of gasoline from natural gas, and ten billion barrels of gasoline from known crude-petroleum reserves. Costs are the limiting factor at present and range from 9 to 18 cents per gallon for Fischer-Tropsch gasoline from coal, 10 to 15 cents per gallon for gasoline from oil shale, 8 to 12 cents per gallon for gasoline from tar sands, 5 to 9 cents per gallon for gasoline from natural gas, against a present cost of 5 cents per gallon for gasoline from crude petroleum. The investment required per barrel of gasoline per day's capacity is also higher, being from $2,800 to $10,000 for the Fischer-Tropsch process compared to $1,400 to $2,800 for the modern refinery using crude petroleum.

Measured by its accomplishments in both peace and war, the American oil industry has much to be proud of. It is a great industry. Because it touches everyone in some manner or other, the adequacy of its resources—crude oil and natural gas—in terms of national needs is rightfully a subject of public concern. This volume, written by top-notch executives of this industry, many of whom are geologically trained, should do much to dispel some of the popular misconceptions concerning the nature of our oil resources.

A. I. LEVORSEN

Introduction à la lecture des cartes géologiques. By ANTOINE BONTE, Assistant de Géologie à la Faculté des Sciences de Besançon. Paris: Masson et Cie, 1945. Pp. 239; figs. 103; pls. 8.

Introduction à la lecture des cartes géologiques is a book specifically on the subject of the geologic map. As M. Eugene Raguin states in the Preface, "The map is together the goal and the instrument of geologic studies." With this in view Antoine Bonte proceeds to show the problems connected with mapping and how it should be done. At the same time he indicates the use of the factual record as it appears on the map as an instrument in further work. Certain aspects of general geology, topographic maps, structure, and stratigraphy as they pertain to a geologic map are also discussed and well illustrated.

Texts on general geology, stratigraphy, structure, and descriptive geometry cover, in part, the material in this book. Usually, however, such books do not present the geological-map aspect of these subjects. In this book Bonte has concentrated on the geological map and has done an outstanding piece of work, which is both stimulating and informative.

A summary of the Table of Contents is as follows:

Chapter i, "The Fundamental Principles of Geology": Stratigraphy, with emphasis on the principles of superposition and continuity.

Chapter ii, "Notes on Tectonics": Types of faults, folds, and nappes.

Chapter iii, "Topographic Maps": Contour and hachure.

Chapter iv, "Geological Maps": French.

Chapter v, "Principal Types of Structure": Some theoretical considerations in connection with them.

Chapter vi, "Instructions on the Geologic Maps": Profiles and geological cross sections and some geomorphology.

Chapter vii, "Geological Mapping": What is needed, and how geological mapping should be done.

Chapter viii, "Graphic Constructions Relative to Geologic Maps": Various applications of descriptive geometry to the problems of geological mapping; exercises.

Annexes A to G: Charts and tables of various types, such as conversion tables, the French stratigraphic column, and keys to French geological maps.

The first two chapters deal interestingly with some of the basic fundamentals of geology, but this material is treated only as it applies to the geological map. The nomenclature of geologic structures is given and illustrated along with a very useful description of nappe structure.

Chapter iii is very informative, discussing, in addition to the contour map, the hachure map, which is an unfamiliar type in the United States. The explanation of hachure-map construction is both interesting and useful.

Chapter iv is valuable for anyone wishing to use French geological maps, as it describes the various types of geological maps available. Anyone having a special interest in a particular section of France may obtain maps of this area by consulting Annexes F and G at the back of the book.

In chapter v, on structure, the author considers the outcrop of beds. The chapter is introduced by examining the appearance of the outcrop of beds having a variety of dips on a fixed slope. This section is well illustrated with block diagrams. The relation between width of outcrop and thickness of bed is also discussed, with numerous formulas to fit all cases. The last part of the chapter is devoted to the characteristic pattern of geologic structures on a map and some of the interpretation and errors of interpretation.

Chapter vi could really be divided into two chapters; the first part deals with general material on the geological map, such as profiles and the construction of geologic cross sections; the last part describes the French stratigraphic column and how some of these formations appear in the field.

The last two chapters would probably be of greatest interest to American geologists. In chapter vii the subject of geological mapping is presented in a different light. All the necessary precautions and "do's" and "don't's" as well as the general procedures are discussed in a refreshing manner. Some of the techniques described are more European than American, but the fundamentals are the same.

The final chapter, viii, on graphic construction, is an excellent discussion. Basic methods of gaining more information from already known facts are diagramed and very well presented. A method of drawing contours on a horizon is described, as well as of getting attitudes of a bed from various points on the same horizon; determining true dip from apparent dip; and various other problems of that nature. Diagrams are a feature of the book, and in the section dealing with geometric constructions there are many useful examples. The last ten pages of the chapter are exercises in the application of the principles presented in the book. There are only four exercises, but each one amounts to more than one simple problem. Two of the problems are completing of a map from known data and making a geologic map from two cross sections. These are excellent problems for anyone.

Bonte felt that there was no work in French which covered this subject and that such a book was needed. It is probably equally true that there is no work in English which covers the subject in the same manner as he has. This book was written from lecture notes; and, in so doing, the author preserved an understanding for the problems of a student. This has produced an excellent and very readable book.

GEORGE J. BELLEMIN

The Submarine Relief off the Norwegian Coast with Bathymetrical Map in Seven Sheets of the Norwegian Coastal Waters and Adjoining Areas. By OLAF HOLTEDAHL. Oslo: Det Norske Videnskaps-Akademi, 1940. Pp. 43; figs. 7; pls. in pocket 6.

The report is essentially a summary description and interpretation of seven colored bathymetrical maps which portray submarine features along the entire Norwegian coast. It includes the following chapters: "Introduction,"

"Description of Some Relief Features as Seen in the Maps," "Notes on the Dredgings Carried Out on Board M/S 'Gunnerus' 1931 and 1932 outside the Romsdalsfjord District," "General Remarks," and "References." The maps, based on the official Norwegian charts, use Mercator's projection and vary in scale from about 1:400,000 to about 1:600,000. The contour-line interval is 10 m. and the color interval 100 m. All data from previously printed official charts and other maps, as well as unpublished recent soundings, have been compiled for these charts. The plates include a composite map of the submarine relief showing the major physiographic features, maps of local areas, and numerous submarine profiles.

The submarine coastal area is dominated by the broad Norwegian shelf (50–200 km. wide), which is regarded as part of the extensive shelf areas of the North Sea to the southwest and Barents Sea to the northeast. Two major tectonic lines outline the shelf: (1) "a zone of dislocation of first order" at the outer margin of the shelf which separates the Eurasiatic land block from the ocean basin and (2) "a zone of fracture of more local character," which separates the Scandinavian land mass from the shelf region. The latter zone is described in detail and is revealed on the maps by linear submarine scarps and troughs margining either the strand flat or the present coast. The scarps in particular have regular, unbroken arcuate trends and are distinct from the pattern of intersecting fractures in the adjoining land mass. They are considered marginal fracture lines of Cenozoic age along which the entire Scandinavian block was uplifted and tilted eastward. This concept of recent uplift, previously advanced by the author along with Nansen, Sederholm, G. De Geer, and others, is substituted for the older view that the Scandinavian mountain range is simply a relict of the higher ancient Caledonian range. In a previous publication (1929) the author suggested that the marginal scarps might mark the contact between crystalline and less resistant younger (Mesozoic?) rocks. The dredgings, however, give no positive evidence, as they indicate glacial bottom materials composed largely of crystalline rocks of Norwegian types.

Except for the Norwegian Channel off the south coast, which is considered a modified tectonic depression, the secondary relief features of the shelf are largely glacial in origin and include: submerged fiord mouths and thresholds, ice-scoured depressions, and moraines. The existence of lobate shelf glaciers is indicated by the moraines, and comparison is made with the Pleistocene glacier tongues along the eastern side of the Andes. The fiords are believed to be mainly the product of ice erosion rather than of faulting. In addition to the submerged glacial features, shallow transverse channels, which in many places can be related to major valleys on the mainland, are interpreted as fluvial and glaciofluvial drainage lines.

Repeated oscillations of sea-level are implied by the shelf features, and three possible submarine levels indicating lower relative sea-levels are noted: (1) 260–70 m., evidenced by abrasion platforms, fiord thresholds, and the flat central part of the Norwegian Channel; (2) 300–400 m., indicated by outer shelf moraines; and (3) below 400 m., suggested by the possible fluvial or glaciofluvial alluvial plain along the northern part of the Norwegian Channel.

It is regrettable that the author's original plan of including a treatment of the geomorphology of the adjoining mainland had to be postponed. This is important for an understanding of the submarine relief, and it is hoped that its appearance will not be long delayed.

LELAND HORBERG

On Major Tectonic Forms of China. By T. K. HUANG. ("Geological Memoirs," Ser. A, No. 20.) Pehpei, Chungking: National Geological Survey of China, 1945. Pp. 165; pls. 8.

Asia, the greatest of continents, probably can add more to our understanding of the earth's deformative history than any other continent. Central and eastern Asia are particularly rich in tectonic features of global importance, as yet rather spottily known. Very welcome, therefore, is this effort to fit together the major tectonic forms of China in a unified treatment which reaches as far as the Karakorams and Lake Baikal.

In his introduction, which is essentially a genetic classification following Stille and Argand, the author discusses continents and geosynclines, orogenic and epeirogenic movements, foundation folds, and, finally, orogenic structures grouped in four principal categories—*Deckengebirge*, *Faltengebirge*, *Bruchfaltengebirge*, and *Blockgebirge*. He presents China's orogenic cycles since the Cambrian as the Caledonian; the Variscan, comprising three subcycles from late Devonian to the end of the Permian; Indosinian (late Trias to Lias); Yenshanian in

three phases (late Jurassic to end of Cretaceous); and the Himalayan in three phases corresponding to those of the Alps.

In the light of these classifications by structures and by time, successive chapters describe the pre-Cambrian massifs, the Caledonides, Variscides, Indosinides, etc. Each comprises many individual geographic units, which are seemingly about as adequately treated in concise form as is possible at present. Following this analysis of the major tectonic features of China and their relationships, chapter ix explains their history and successive development through the geologic ages.

Progressing to larger relationships, chapter x, "The Geotectonic Make-up of China," develops various principles of continental evolution substantiated by China's contributions. The discussion includes: (1) areas of crustal weakness versus areas of crustal strength; (2) the order and intensity of folding in a given orogenic cycle; (3) consolidation, migration of geosynclines, and growth of the continent; (4) arcuate fold systems and their origin; and (5) *Leitlinien* and major tectonic types The southward migration of geosynclines, with their subsequent foldings outward from the old Angaraland massif is our most declared example of the growth of a continent. Subsequent chapters treat the igneous activity and metallogenesis by epochs and by provinces.

To anyone interested either in the regional tectonics of China or in the part which southeastern Asia has played in the general tectonic development of the globe, this report can be strongly recommended.

R. T. C.

India, Part I: *Physical Basis of Geography of India*. By H. L. CHHIBBER. Benares: Nand Kishore & Bros.; 1945. Pp. 282; figs. 19; pls. 10. Rs. 5; sh. 9.

Recognizing that the physical basis of the geography of a country is necessary for a proper grasp of its geography as a whole, the author has prepared this short treatise to meet a generally felt need. The chapters of the first half of the book present in succession the physical divisions, mountains, plateaus, rivers, lakes, glaciers, earthquakes, coastline, volcanoes, and the hot springs and mineral waters of India. They are largely descriptive and informative, presenting briefly the most important facts. Some readers, however, will be surprised by the statement on page 14: "This mountain [K2 28,250 ft., generally considered the world's second highest peak] was ascended by the Duke of Abruzzi in June, 1909, and it represents the highest altitude ever climbed by a human being."

The geologic rock systems of India are described in concise fashion for the reader who wishes a general survey. The "Making of India" explains the dynamic history, which is amplified by a later chapter on the "Structure of India." Soils, soil erosion, and fertilizers are given relatively full treatment, being particularly vital to a populous country like India.

R. T. C.

German-English Science Dictionary. By LOUIS DE VRIES. 2d ed. New York and London: McGraw-Hill Book Co., 1946. Pp. xiv+558. $4.50.

The first edition of this dictionary, prepared with the collaboration of members of the graduate faculty of Iowa State College, was reviewed in the *Journal of Geology*, Vol. XLVIII (1940), p. 1022. The new revised edition has been enlarged, particularly in the fields of chemistry, biology, agriculture, and forestry.

R. T. C.

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DRAINAGE DIVERSION IN THE NORTHERN ROCKY MOUNTAINS OF EAST-CENTRAL IDAHO[1]

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Cornell University

ABSTRACT

The upper Salmon River, many of its tributaries, and other streams draining the northern Rocky Mountains of east-central Idaho flow for many miles in a northeasterly direction and then abruptly change their courses. The Salmon River makes an elbow turn and virtually reverses its direction of flow to the west side of the state, joined below by important barbed tributaries and above by consequent streams which occupy northwestward-trending structural basins. Other northeast-flowing streams enter structural basins and are directed southeast to the Snake River Plain.

Headwater trends and aligned wind gaps across block mountains are indicative of an early drainage system that extended far to the northeast, probably to the Missouri River. That piracy and large-scale faulting have had an important role in disrupting this early drainage is indicated by barbed tributaries, elbows of capture, wind gaps on tilted block mountains, and the presence of consequent streams in fault-made valleys.

Diversion of Salmon River to the west side of the state apparently resulted from capture by headward erosion of a vigorous stream from the west as a consequence of crustal disturbances during late Tertiary and early Quaternary time. Diversion of other streams into northwest-southeast structural basins between block-faulted mountains probably took place at about the same time. By reason of these diversions the Continental Divide has been shifted about 100 miles east of its location in late Tertiary time.

INTRODUCTION

The drainage in the northern Rocky Mountains of east-central Idaho has a peculiar pattern indicative of a complicated and an eventful development. The Salmon River and some of its important tributaries trend across the area in a northeasterly direction almost to the Montana line and Continental Divide where the main river then makes a sharp elbow turn and virtually reverses its direction of flow to cross to the west side of the state, joined in its westward course by tributaries which are barbed and which, until they enter the Salmon, are parallel to the headwater streams (Fig. 1). Other streams, like Big Lost River, also trend in a northeasterly direction, but after some miles they abruptly change their courses and flow as consequent streams in long, intermontane basins either northwest to join the Salmon River or southeast to disappear on reaching the Snake River Plain. Wind gaps aligned with the headwater streams and extending across the mountains bordering the basins indicate that the drainage formerly extended much farther to the northeast than it does at present.

The thesis here presented is that the streams of northeast trend represent the headwaters of a previously established

[1] Presented at the Chicago meeting of the Geological Society of America, December, 1946.

Linlithgow Library.

drainage system that extended east of the Continental Divide, perhaps to the Missouri River, and that the existing peculiar stream pattern is the consequence of piracy and large-scale faulting that disrupted the earlier drainage and caused the Continental Divide to be shifted about 100 miles east of its former location.

TOPOGRAPHIC SETTING

As constituted locally, the northern Rocky Mountains are composed in part of dissected uplands and in part of block mountains and intermontane basins like those in Utah and Nevada (Pl. I).

Dissected uplands.—The dissected uplands make up about two-thirds of the area under discussion, composing the

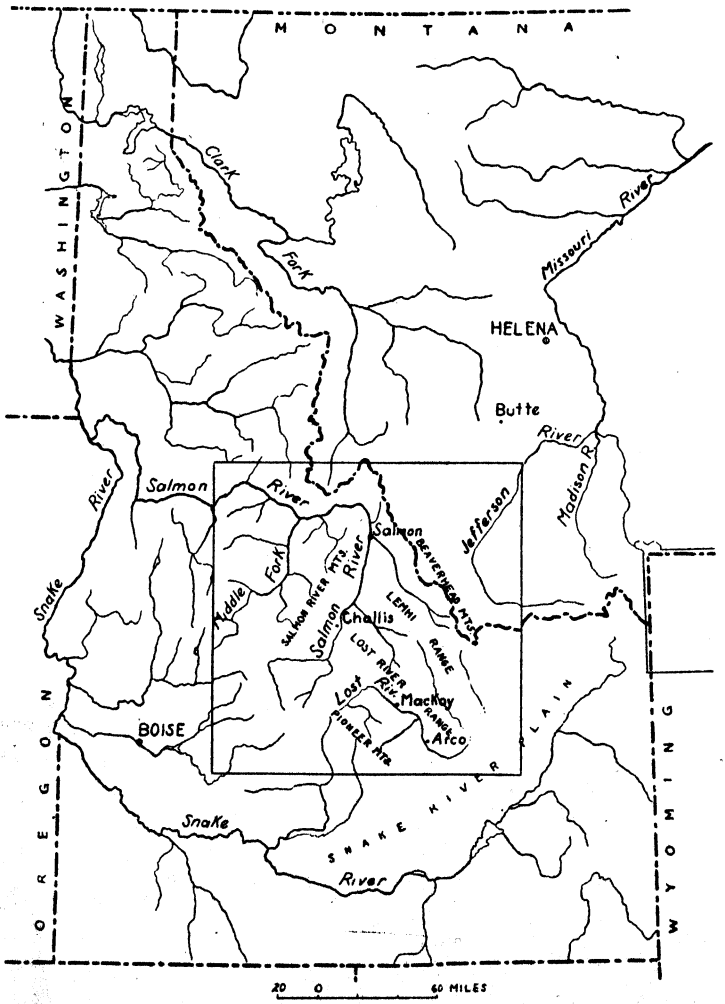
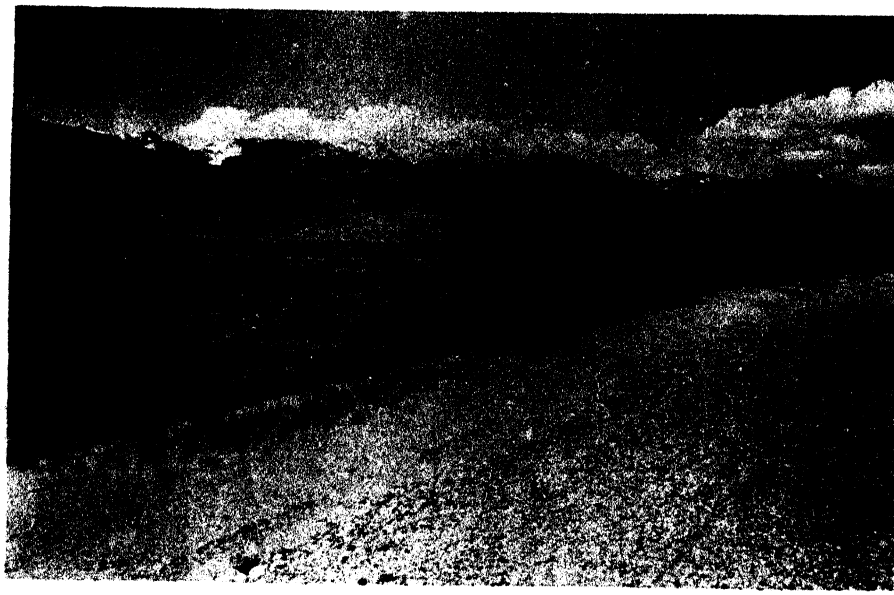


FIG. 1.—Index map showing location of diverted drainage systems in east-central Idaho

PLATE I

Relief map of east-central Idaho showing the northern Rocky Mountains composed locally of dissected uplands and isolated basins and ranges.

A



B

A, Dissected uplands comprising the Salmon River Mountains in the vicinity of Panther Creek. *B*, Desert basin of Birch Creek with the tilted back slope of Lemhi Range in the background.

mountains in the northern and western part. Although the uplands form essentially a continuous mountain group, they have been divided into poorly delimited units which have been named from the drainage systems that dissect them or from prominent physiographic or other features that locally may distinguish them. These units include: the Sawtooth Mountains at the head of Salmon River; the Salmon River Mountains, north and west of Salmon River, mountains which extend from the headwaters of the river in the Sawtooth Mountains to its elbow turn near the Montana line; the White Cloud Peaks, which are near the west edge of an unnamed group between the Sawtooth Mountains and the eastern border of the upland group; the Boulder Mountains southeast of the Sawtooth Mountains and White Cloud Peaks; the Pioneer Mountains between the Boulder Mountains and the Snake River Plain; and the White Knob Mountains between the Pioneer Mountains and the eastern border of the uplands (Fig. 2).

As pointed out by J. B. Umpleby,[2] the dissected uplands composing the largest unit, the Salmon River Mountains, show a marked accordance of ridge levels and bear the remnants of an old erosion surface that has been deeply and intricately trenched by the present drainage system (Pl. II, *A*). This erosion surface may also be recognized in the other mountain units, though not everywhere so well preserved as in the Salmon River Mountains. Although essentially flat over broad areas, the surface shows the effect of differential uplift and locally may form the flanks and buried floors of minor intermontane basins and the crests and back slopes of none-too-prominent mountain blocks. Over much of the Salmon River Mountains the upland surface stands 8,000 to 10,000 feet above sea-level. In some of the other units, ridges rise above 10,000 feet and some peaks above 11,000 feet—the highest, Hyndman Peak in the Pioneer Mountains, reaching 12,078 feet. Despite some topographic discontinuities, however, the upland areas retain all the essential characteristics of a greatly dissected peneplain or erosion surface which has been incised to depths of as much as 5,000 feet by some of the major streams.

Basin and range area.—The basin-and-range part constitutes the southeast third of the area, though actually this part extends well up the eastern border, almost to the elbow turn of Salmon River (Pl. I). It is separated from the upland group by a great intermontane trench that reaches from the Snake River Plain northwest to the Salmon River, a distance of 90 miles, and from that point on by the Salmon River. In addition to this bordering trench, the area embraces three parallel mountain ranges, two of them 90 miles long, the third about 145 miles long, and two intervening trenches 100 and 110 miles long, all extending northwest from the Snake River Plain to the Salmon River (Pl. I). The ranges are known, respectively, from southwest to northeast as the "Lost River Range," the "Lemhi Range," and the "Beaverhead Mountains," the latter forming a part of the Bitterroot Range and carrying on its crest the Continental Divide. The trenches between the ranges and between the Lost River Range and the dissected uplands have been named from the streams that drain them; and, since drainage within them is both northwest and southeast from low, about centrally located valley-floor divides, a part of each trench is known by one name, the remainder, by another. These parts are

[2] "Geology and Ore Deposits of Lemhi County, Idaho," *U.S. Geol. Surv., Bull. 528* (1913), pp. 22–24.

basins of the desert type (Pl. II, *B*), but instead of being designated as basins they are called "valleys," and those separating the Lost River Range from the dissected uplands are referred to as the Warm Spring and Big Lost River valleys, those between the Lost River and Lemhi ranges as the Pahsimeroi and Little Lost River valleys, and those between the Lemhi Range and the Beaverhead Mountains as the Lemhi and Birch Creek valleys (Pl. I). Warm Spring, Pahsimeroi, and Lemhi valleys drain northwest to the Salmon River; the others, to the Snake River Plain.

Topographically the mountain ranges are much alike. Each of them presents an abrupt scarplike front to the broad basins along their southwest borders and less evenly faced slopes on their north-

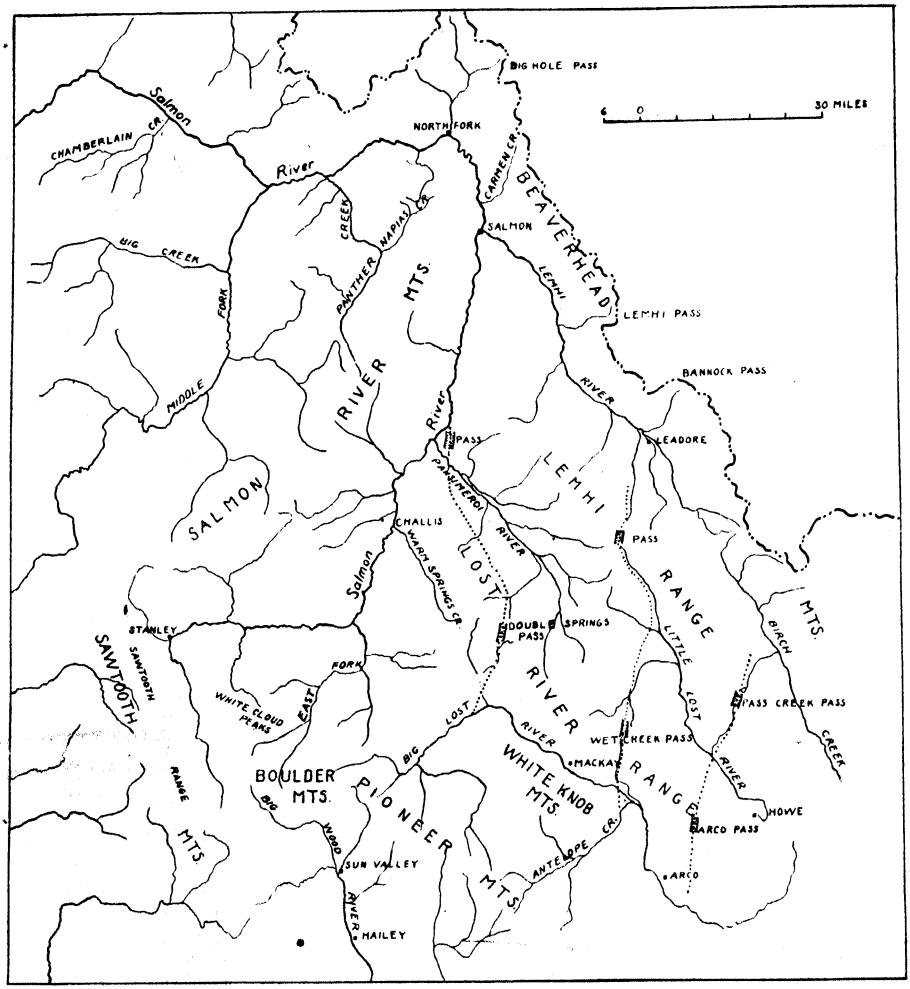


FIG. 2.—Sketch map showing location of mountains, streams, and wind gaps (passes) and former drainage courses (*dotted*).

east sides. In each the crest is close to the southwest border, and each therefore resembles an asymmetrical, steeply tilted block—the southwest side representing the front of the block, the northeast side, the back (Fig. 3). On the southwest sides the regularity of the scarplike faces has been interrupted somewhat by sharply incised V-shaped canyons ending at the bases of the ranges and opening upon extensive alluvial fans (Pl. III, *A*). On the back slopes of both the Lost River and the Lemhi ranges are tilted remnants of an erosion surface like that which distinguishes the upland area. Near the north end of the Lemhi Range the surface becomes broader and flatter and matches the one in the Salmon River Mountains across the Salmon River. The summit erosion surface is exceptionally well preserved in parts of the Beaverhead Mountains, in places being notably flat or gently undulating and little dissected.[3] Across each of the ranges are relatively low natural passes (wind gaps) to which reference will be made later. Except for these passes, the ranges are little notched and present bold, precipitous fronts reminiscent of the Wasatch Range in Utah.

These ranges are among the highest in the state, all rising above 11,000 feet and two of them above 12,000 feet. Lost River Range carries the highest point in the state, Borah Peak at 12,655 feet, which towers 6,000 feet above the bordering basins (Fig. 3). When connected, the high points of these ranges are along an axial line that is parallel to and about 50 miles from the edge of the Snake River Plain. This axis, one of upwarp, parallels the Snake River Plain across the state, passing through Hyndman Peak in the Pioneer Mountains and through other high mountains to the west and northwest. On either side of the line the mountains descend gradually to lower levels, those on the south side to the level of the Snake River Plain.

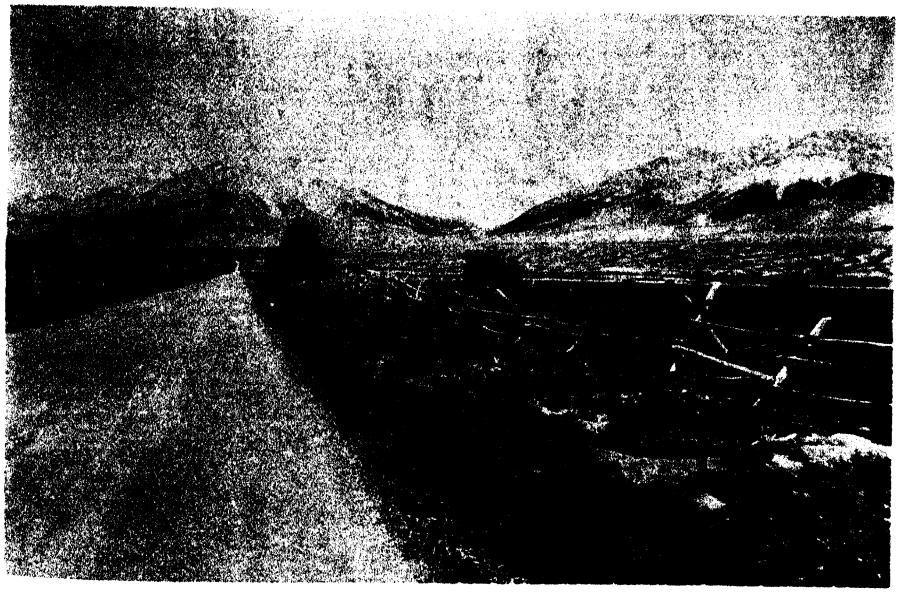
The intermontane trenches are singularly straight and remarkably uniform in width (Pl. I and Pl. II, *B*). Their northeast borders are especially straight and regular, their southwest borders, where they touch against the back slope of the ranges, being less so. The trenches containing the Birch Creek and Lemhi basins and the Little Lost River and Pahsimeroi basins average 5 to 8 miles wide, although at its head the Pahsimeroi basin is but 2 miles wide and in its lower part as much as 10 miles wide. The trench containing Big Lost River and Warm Spring basins is narrower, mostly 3 to 5 miles wide. All are surfaced by great alluvial slopes extending out from the base of the mountain fronts. As indicated earlier, these trenches drain in opposite directions from low invalley divides. The divide which separates the drainage of Birch Creek from that of the Lemhi River is almost imperceptible, and for about 8 miles has no surface drainage at all. At the divide the trench is as wide as at any place along its course. From about 7,200 feet the Birch Creek basin drops to 5,800 feet where it merges with the Snake River Plain 40 miles to the southeast, and the Lemhi basin descends to 4,000 feet at the Salmon River 70 miles in the opposite direction. The divide between Little Lost River basin and the Pahsimeroi, although restricted by the Donkey Hills, is still several miles wide and is about as imperceptible as the divide between the

[3] P. J. Shenon, "Geology and Ore Deposits of the Birch Creek District, Idaho," *Idaho Bur. Mines and Geol. Pamph. No. 27* (1928), pp. 3–4; A. L. Anderson and W. R. Wagner, "Lead-Zinc-Copper Deposits of the Birch Creek District, Clark and Lemhi Counties, Idaho," *Idaho Bur. Mines and Geol. Pamph. No. 70* (1944), p. 4.

FIG. 3.—Topographic map (Borah Peak quadrangle) showing a part of the asymmetrically block-tilted Lost River Range, its southwestward-facing frontal escarpment, and the wind gap (Double Springs Pass) that crosses the range in alignment with the upper course of Big Lost River.

PLATE III

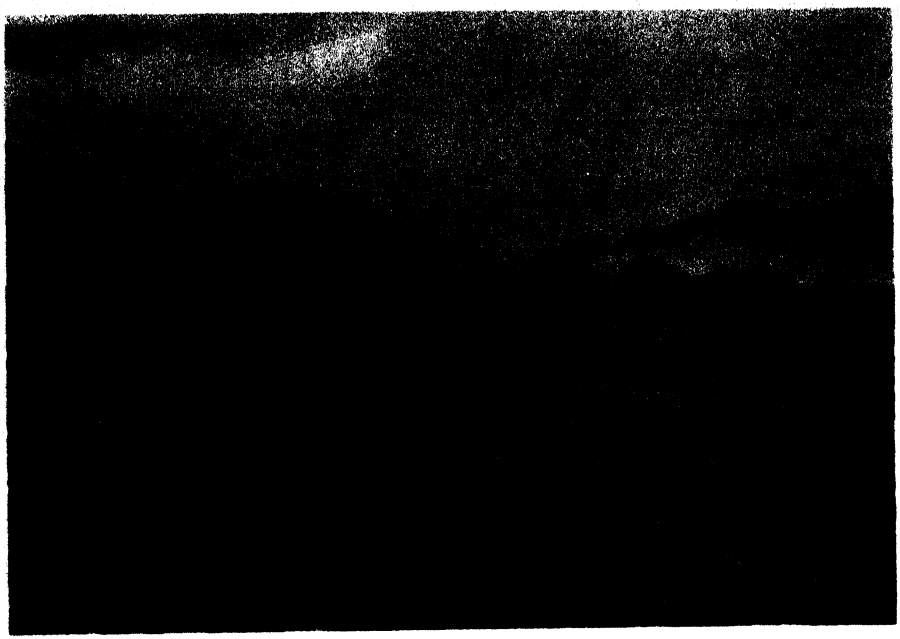
A



B

A, Lost River Range escarpment from upper Warm Spring Valley, showing aligned faceted spurs, sharply incised V-shaped canyons and fronting alluvial fans. *B*, Double Springs Pass (wind gap) which crosses Lost River Range opposite the place at which Big Lost River enters and flows southeast along Big Lost River Valley. The wind gap is interpreted as a former valley of Big Lost River before its defeat and diversion by block faulting.

PLATE IV



A

B

A, Wet Creek Pass (wind gap) which crosses Lost River in line with Antelope Creek is interpreted as a former course of Antelope Creek before its diversion by block faulting. *B*, Arco Pass (wind gap) near the southeast end of Lost River Range as viewed from Little Lost River Valley. The head of the former Arco River, which occupied the wind gap, has been carried below by the Snake River downwarp and covered by lava.

Birch Creek and Lemhi River basins. From the divide at 6,728 feet above sea-level the Little Lost River basin extends 55 miles southeast to the Snake River Plain and the Pahsimeroi River basin 45 miles northwest to the Salmon River. The divide between the Big Lost River and Warm Spring Creek basins is more pronounced, a prong from the upland area extending across to the base of the Lost River Range (Fig. 3). However, the divide at Willow Creek summit is 7,161 feet and therefore no higher than the one in the Birch Creek–Lemhi River trench. From this divide the Warm Spring Creek drainage extends to the Salmon River 30 miles away and the Big Lost River drainage to the Snake River Plain 60 miles to the southeast. Like the high points on the ranges, the basin divides are close to an axial line that parallels the margin of the Snake River Plain and that coincides with that of the upbowed ranges.

PREVIOUS INTERPRETATIONS

The drainage pattern within the uplands has not previously attracted attention, but considerable interest has been focused on the drainage in the eastern basin-and-range area. Umpleby[4] was first to comment on the origin of the basins and the drainage in them. The basins he regarded as ancient valleys carved in an early erosion surface (peneplain) and the streams that carved them as belonging to a southeast drainage system that extended to the Snake River Plain. The long valley trench occupied by Lemhi River and Birch Creek he regarded as occupied by an ancient stream that headed north of the town of Salmon whose headwaters were later captured by the Salmon River and the divide pushed back to its present location by headward erosion of Lemhi River. He later[5] mentioned re-entrant valleys in the Lost River Range and explained them as belonging to the older drainage system and stated that the one near Arco was once probably occupied by Little Lost River until blocked by lavas in the vicinity of Arco Pass, when the river was diverted into its present channel. The valley of Big Lost River he thought could be accounted for, more certainly than any of the other large valleys, by the erosional activity of the stream that now occupies it, though faulting may have been a factor in the development of the valley, as indicated by the triangular faces on the eastern side.[6]

After making a particular study of the drainage in western Montana and eastern Idaho, W. W. Atwood[7] came to the same conclusion as Umpleby concerning the southeast drainage, but he stressed warping and faulting in the development of the valleys (basins).

O. E. Meinzer[8] later called attention to the contrast between the straight, well-defined valley front on the northeast side of the Pahsimeroi Valley and the indefinite border on the southwest side and suggested faulting as the chief means of valley depression.

V. R. D. Kirkham[9] then expressed a

[4] P. 30 of ftn. 2.

[5] "Geology and Ore Deposits of the Mackay Region, Idaho," *U.S. Geol. Surv., Prof. Paper 97* (1917), pp. 19–21.

[6] Pp. 20–21 of ftn. 5.

[7] "The Physiographic Conditions at Butte, Montana, and Bingham, Utah, When the Copper Ores of These Districts Were Enriched," *Econ. Geol.*, Vol. XI (1916), pp. 706–21.

[8] "Ground Water in Pahsimeroi Valley, Idaho," *Idaho Bur. Mines and Geol. Pamph. No. 9* (1924), p. 16.

[9] "A Geologic Reconnaissance of Clark and Jefferson and Parts of Butte, Custer, Fremont, Lemhi, and Madison Counties, Idaho," *Idaho Bur. Mines and Geol. Pamph. No. 19* (1927), p. 11.

view that the depressions were undoubtedly structural and antecedent to the streams that now occupy them but that the relations of several re-entrant valleys, which have a decided northeast trend, indicated that the ancient drainage flowed northward and not to the south as proposed by Umpleby and Atwood.

Soon after, Shenon[10] pointed out that block faulting, in places with a vertical displacement of 3,500 feet, defines the wide valley in which Birch Creek flows and that the relatively straight, clean-cut valley wall with well-defined faceted spurs extends into the Lemhi basin, indicating extensive development of the faulting. Having been over much of the area in question and after a careful review of the literature, he accepted the view of Umpleby and Atwood for a southeastward-flowing drainage system, despite the contradictory evidence pointed out by Kirkham, which, he stated, the broader evidence apparently failed to support.

C. P. Ross[11] has since suggested an origin for the Warm Spring Creek basin and the northwest end of the Lost River Range other than by faulting, but E. M. Baldwin,[12] working in the same region, has concluded that Lost River Range is a tilted fault block and that the bordering basins are fault-controlled.

More recently, the writer[13] has confirmed the fault-block origin of Birch Creek Valley and has further evidence to back up an earlier publication[14] that the three ranges have been produced by block faulting and that the "valleys," so called, are structural basins. In neither of these reports were the drainage problems discussed. Evidence is now to be presented to show that the former drainage was northeast rather than southeast and that capture of Salmon River and block faulting have succeeded in beheading the drainage and directing it into its present channels.

EVIDENCE OF EARLY NORTHEAST DRAINAGE

Evidence of an early northeast drainage system is preserved in the present headwater streams, in the northeast alignment of wind gaps across the block ranges as prolongations of these streams, and in the distribution pattern of Tertiary volcanics.

Headwater drainage.—The evidence afforded by headwater streams is to be found largely in the dissected upland area where the drainage pattern has been least modified by differential crustal movements. The upper Salmon River itself offers a most striking clue to the direction of the early drainage (Fig. 2). For the first 40 miles of its course it flows in a northwesterly direction along virtually the full length of Stanley Basin, an intermontane fault-made basin extending along the base of the Sawtooth Range, a tilted-block segment of the Sawtooth Mountains (Pl. I). Near the settlement of Stanley it leaves the wide alluvial-floored depression and flows eastward in a deep, steep-sided valley that widens somewhat downstream. Near its junction with its East Fork

[10] P. 4 of ftn. 3.

[11] "Erosion in the Lost River Range, Idaho," abstract in *Jour. Wash. Acad. Sci.*, Vol. XXVIII (1938), p. 415.

[12] "Structure and Stratigraphy of the Northern Half of Lost River Range, Idaho" (Ph.D. dissertation, Cornell University, 1943).

[13] P. 13 of ftn. 3.

[14] A. L. Anderson, "A Preliminary Report on Recent Block Faulting in Idaho," *Northwest Science*, Vol. VIII (1934), pp. 17–28.

(Fig. 2) it changes its course from east to northeast, and its valley continues to widen. Just above the town of Challis it debouches into the comparatively broad expanse of Warm Spring Creek basin (Pl. I), skirts the lower end of the basin, and then a short distance below Challis enters the deep narrow canyon that separates the Lost River Range from the Salmon River Mountains. Emerging from the canyon, the river, which continues to flow in a north-northeasterly direction, skirts the lower end of the broad Pahsimeroi River basin and then enters the deep 30-mile-long canyon that cuts across the upland area and separates the Lemhi Range from the Salmon River Mountains. About 10 miles above (south of) the town of Salmon the river emerges from the canyon into the broad basin of Lemhi River and flows in a more northerly direction along the lower northwest edge of the basin to a point about 10 miles below (north of) Salmon, where it enters a canyon from which it never again emerges. As it enters this canyon, it swings from north to northwest and then at the settlement of North Fork makes a right-angle turn and from there crosses to the west side of the state.

The main tributaries of the upper Salmon also conform with it in trend. East Fork, which rises in the White Cloud Peaks and Boulder Mountains area, flows in a northeasterly direction aligned with the Salmon River at and below the point about 35 miles below Stanley where the Salmon changes its course from east to northeast (Fig. 2). The Middle Fork of Salmon River, which has its head at or near the northwest end of the Sawtooth Range (Fig. 2), flows in a northeasterly direction to its junction with the main Salmon about 35 miles below North Fork. Its course is almost exactly parallel to that of East Fork and Salmon River below the East Fork junction. Its principal tributary, Big Creek, joins from the west and from its head to its mouth flows nearly due east. Chamberlain Creek is another tributary of the main Salmon that flows in an east-northeasterly direction. It joins the main Salmon a few miles below the mouth of Middle Fork (Fig. 2). Panther Creek is another important tributary, which, except for its lower few miles, flows in a northeasterly direction. At the point where its course changes from northeast to northwest it is joined by Napias Creek, which, though flowing southwest, is aligned with the upper course of Panther Creek.

The northeast trend of the headwater drainage is also shown by streams other than the Salmon and its main tributaries. Big Lost River, which rises in the Boulder and Pioneer Mountains, flows northeast until it reaches the broad floor of Big Lost River Valley (Pl. I and Fig. 2) where it turns sharply southeast and follows the front of Lost River Range to the Snake River Plain (Pl. I). Antelope Creek, a tributary of Big Lost River, rises in the Pioneer Mountains and flows northeast to its junction with Big Lost River in Big Lost River Valley. Even the head of Pahsimeroi River, which has its source in the Lost River Range, flows obliquely across the back slope of the range (Fig. 3) in a general north-northeasterly direction until it emerges upon the broad floor of Pahsimeroi Valley (basin). Other streams that head on the northeast sides of both the Lost River and Lemhi ranges also trend obliquely across instead of directly down the back slopes, thus conforming with the northeast drainage trends. As they debouch upon the broad plains at the foot of the ranges, they are directed

either northwest or southeast along the basins.

Wind gaps.—The existence of wind gaps or passes, which are aligned in a northeast-southwest direction across the block ranges, is also evidence of an earlier northeast drainage. One of the most impressive of these wind gaps is Double Springs Pass (Fig. 3) which extends across Lost River Range in a north-northeasterly direction opposite the point at which Big Lost River emerges from the uplands into Big Lost River Valley (basin). This pass (Pl. III, *B*) is cut 3,000 to 4,000 feet into the range, its high point of 8,318 feet being in line with the crest of the range and close to its southwest border. The pass constitutes a transverse valley tilted and flaring toward the Pahsimeroi Valley basin. The old valley course is partly choked with morainal debris and outwash from the bordering slopes, and earlier stream gravels are not exposed. The valley is not occupied by permanent streams, and much of the drainage today is subsurface. This old valley crossing is in direct line with Big Lost River as it emerges from the upland area. In crossing the range the old valley swings to the north-northeast and shows alignment with a wind gap that cuts across a corner of Lemhi Range very close to the lower end of Pahsimeroi Valley (Fig. 2). This second wind gap is about 1,000 feet above the Pahsimeroi and Salmon River Valley floors and appears as a deep notch connecting the Pahsimeroi Valley basin with the Snake River Canyon. If Big Lost River were extended through Double Springs Pass and this second wind gap, it would make a perfectly normal tributary of the Salmon River.

Wet Creek Pass (Fig. 2 and Pl. IV, *A*) is another wind gap that forms a deep notch across Lost River Range, in this case essentially opposite the mouth of Antelope Creek. It has been described by Kirkham[15] as a re-entrant valley and has been used by him as evidence of an old northeastward drainage course. This wind gap repeats the essential features of Double Springs Pass. It cuts deeply into the range, its floor being about 7,500 feet above sea-level, and forms a transverse valley tilted northeast. It is floored with gravel washed down from bordering slopes, but gravel caps on ridges carved by Wet Creek in the floor of the valley may consist of old stream gravels. This valley trends north-northeast and is aligned with Antelope Creek as Antelope Creek approaches the edge of the upland area. It is also in line with a long and deep wind gap that extends across the Lemhi Range at the head of Little Lost River (Fig. 2). This second wind gap may be seen to good advantage from the Lemhi Valley near the town of Leadore. Whether Bannock Pass in the Beaverhead Mountains is a prolongation of these wind gaps has not been determined.

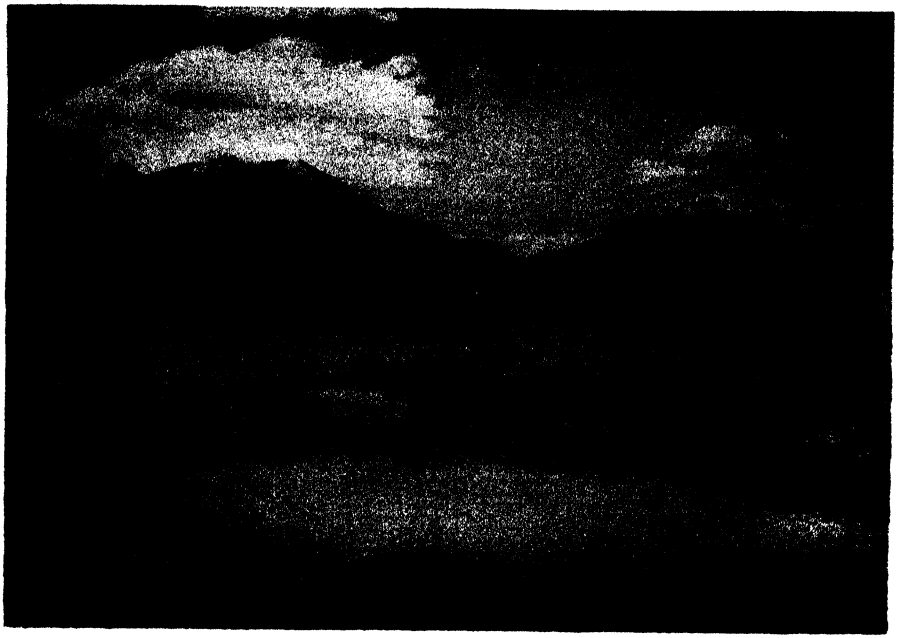
Arco Pass (Fig. 2 and Pl. IV, *B*) close to the southeast end of Lost River Range also attracted the attention of both Umpleby[16] and Kirkham[17] and prompted each to regard it as a former drainage course—regarded by Umpleby as a southwest drainage course, by Kirkham as a northeast drainage course. This pass, or "re-entrant valley" as called by Kirkham, extends in a north-northeasterly direction; but, unlike the other passes along the Lost River Range, it is not opposite any present streams. Instead, it opens upon the Snake River Plain, standing about 1,500 feet above the surface of the plain. The pass is fairly broad

[15] Pp. 10–11 of ftn. 9.

[16] P. 19 of ftn. 5.

[17] Pp. 10–11 of ftn. 9.

PLATE V



Pass Creek Pass (wind gap) which extends across Lemhi Range in line with Arco Pass and was formerly occupied by the defeated and no longer existent Arco River.

(Pl. IV, *B*); and, like the other passes, its floor is tilted northeast and merges with the broad expanse of Little Lost River Valley. In line with it across the Lemhi Range is Pass Creek Pass (Fig. 2 and Pl. V), another deep transverse valley notch of northeast trend. This pass stands out conspicuously from both sides of the Lemhi Range, being stranded 1,000 feet or more above the basins at either end. The head of the stream that carved these now wind-gap valleys has apparently disappeared beneath the surface of the Snake River Plain, but the record of its former northeast trend is indicated by the aligned wind gaps across the Lost River and Lemhi ranges. Its course beyond the Lemhi Range has not been followed.

There are other minor wind gaps that cross the Lost River and Lemhi ranges, and these also afford evidence of a former northeast drainage system. There are also a number of passes that extend across the Beaverhead Mountains, but there was no opportunity in the field to match these with those that cross the Lemhi Range. From southeast to northwest these are the Bannock Pass at 7,672 feet, formerly utilized by the Gilmore and Pittsburgh Railroad; Lemhi Pass at 8,000 feet at the head of Agency Creek; Big Hole Pass at 7,236 feet and just beyond the point where Salmon River turns west; and Gibbonsville Pass at 6,951 feet at the head of North Fork of Salmon River.

Distribution pattern of Tertiary volcanics.—Some of the belts of Tertiary volcanics also indicate an old northeast drainage system. As pointed out by Umpleby[18] and by C. P. Ross,[19] the Tertiary volcanics blanketed a region of considerable relief, in part with as much relief as that which exists today. The volcanics filled the old valleys, and, although the volcanics have been stripped from considerable portions of the region by subsequent erosion, many of the old valley fills still remain and are exposed as belts which may be traced across the area, commonly in a northeast and north direction. Some of the present valleys as well as some of the older wind-gap valleys are directed along these old valley fills. The Salmon River, at least from Challis to and beyond Salmon, is along a belt of volcanics, only remnants of which, in places, remain in upper canyon walls. Another belt extends through Arco Pass, parallel in trend with the present high-level valley. The pre-volcanic valley was considerably broader but little deeper than the re-entrant valley now carved along it. There is another belt of volcanics, through or alongside Wet Creek Pass, which is aligned with patches of volcanics along the south tip of Donkey Hills at the head of Little Lost River Valley and across in the Lemhi Range extending toward the head of Little Lost River. Bannock Pass in the Beaverhead Mountains also is carved in an older volcanic-filled valley. Belts of volcanics extending in a northeasterly direction from Fish Creek and Muldoon to Mackay and vicinity[20] afford additional evidence of an early northeast drainage trend.

EVIDENCE OF DRAINAGE DIVERSION

Diversion of the early northeast drainage into its present form is indicated by barbed tributaries and elbows of capture and by abrupt changes in direction along

[18] Pp. 20 and 34 of ftn. 5.

[19] "Geology and Ore Deposits of the Bayhorse Region Custer County, Idaho," *U.S. Geol. Surv., Bull. 877* (1937), p. 87.

[20] Pl. I of ftn. 5.

fault-made basins in which the streams now occur as consequents.

Salmon River.—Diversion of Salmon River by capture and its virtual reversal of flow is indicated by its sharp elbow turn a few miles below (north of) Salmon and its many barbed tributaries from that point on (Fig. 2). As it makes the turn, it leaves the broad basin of Lemhi Valley, and its direction, which for many miles both in and on the other side of the basin has been north-northeast, changes to northwest and then at North Fork to sharply west. At the same time it enters a profound canyon extending to the Snake River on the opposite side of the state.

Even as the Salmon River begins its elbow turn, its tributaries become barbed, the first, which is Carmen Creek, being almost directly in line with the upper course of the river but with its flow exactly reversed with respect to that of the Salmon. Umpleby[21] apparently noticed this relationship, which prompted him to conclude that the drainage was south until its capture by a vigorous stream from the west. It is possible that Carmen Creek is the inverted course of the northeasterly flowing Salmon River before its capture.

Downstream almost all the tributaries, except the smaller subsequents, are barbed. Chamberlain Creek and Middle Fork of the Salmon River show this relationship exceptionally well (Fig. 2). Panther Creek, however, enters the Salmon at a normal angle, but the alignment of Napias and Moose Creek valleys with upper Panther Creek (Fig. 2) and certain abnormalities of the Napias and Moose Creek valleys suggest that Panther Creek may have been captured by a subsequent tributary of the Salmon and its former course to the northeast left to Napias and Moose creeks. Both Napias and Moose creeks drain old, relatively shallow valleys aligned with upper Panther Creek and are separated from each other by a low valley-floor divide. In flowing southwest, Napias Creek enters a narrow gorge and cascades abruptly into the present valley of Panther Creek. Moose Creek does likewise in the opposite direction, entering Salmon River as a barbed tributary. Traced northeastward, the old valley course of Panther Creek would enter the drainage area of North Fork of Salmon River and possibly would extend through either Big Hole or Gibbonville passes.

Big Lost River.—Some of the headwater tributaries of Big Lost River are controlled by northwest-southeast structural lines, but the main river crosses structural trends and flows northeast until it reaches the broad basin that extends southeast along the base of Lost River Range. It there makes a sharp elbow turn (Pl. I and Fig. 3) and flows down the basin as a consequent stream. That the basin is fault-developed is evident from its relations to Lost River Range. This range has every characteristic of a tilted fault block with the basin on the downthrow side. Unable to continue its course to the northeast across the rising mountain block, the defeated river was diverted southeast along the structural basin. The stranded valley (wind gap) of Double Springs Pass that crosses the Lost River Range remains as evidence of Lost River's former northeast extension and subsequent disruption.

Antelope Creek.—Antelope Creek repeats all the essential features of Big Lost River, except that, as it enters Big Lost River Valley at the base of Lost River Range, it becomes a tributary of Big Lost River. Its course thus changes

[21] Pp. 29–30 of ftn. 2.

abruptly from northeast to southeast as its waters join those of Big Lost River. That its former course extended far to the northeast is indicated by the aligned transverse valleys (wind gaps) across the Lost River and Lemhi ranges. Antelope Creek is thus a defeated stream.

Basin drainage.—Each intermontane basin is deeply floored with gravelly deposits made by streams from the bordering mountains. Some of the streams emerge from valleys cut obliquely across the back slopes of the ranges, apparently retaining previously established northeast drainage courses, but many flow directly down the back slopes. Drainage within each of the basins is by consequent streams. Warm Spring Creek and the Pahsimeroi and Lemhi rivers thus occupy structural basins which drain to Salmon River; Big and Little Lost rivers and Birch Creek occupy structural basins with outlet to the Snake River Plain.

DATING THE DRAINAGE DIVERSION

Since the old erosion surface that forms the summits of the dissected uplands and the back slopes of the tilted block ranges antedates the drainage diversion, it assists in limiting the time at which stream capture and diversion have taken place. As pointed out by Ross,[22] this surface is younger than the deformation of the Challis volcanics (late Oligocene or early Miocene). Assuming that much of Pliocene time might be necessary to truncate the disturbed volcanic strata and older rocks, the planation may not have been completed until well toward the close of the Tertiary period. Drainage until then and later must have continued in a northeast direction.

Capture and reversal of Salmon River probably were accompanied by or closely followed the late Tertiary orogeny when uplift of the region induced vigorous headward erosion of a westward-flowing stream which finally reached back to the main Salmon and diverted it from its northeast course. Block faulting associated with the same orogeny apparently diverted Big Lost River and other northeastward-trending streams into northwest-southeast structural basins. The diversion was completed be ore Wisconsin glaciation in Pleistocene time, for heads of the streams show glacial sculpture and valleys below, and some of the wind gaps contain morainic deposits and outwash trains. Whether the diversions were as late as the early glaciation which antedated the Wisconsin by more than 1,000 feet of canyon cutting[23] is not known. The depth of the wind gaps indicates considerable downcutting before the streams were defeated by faulting. The drainage changes could have been brought about in early Quaternary time. As pointed out by Ross,[24] the record since the formation of the post-Challis surface has been one of intermittent uplift and vigorous erosion, the details of which are not yet fully known. Because of the excellent preservation of the fault scarps, considerable of the movement that delineated the mountain blocks may have taken place in early Quaternary time.

HISTORY OF DRAINAGE DEVELOPMENT

According to Ross,[25] much of the western part of the area (site of the present

[22] Pp. 89–91 of ftn. 19.

[23] C. P. Ross, "Early Pleistocene Glaciation in Idaho," *U.S. Geol. Surv., Prof. Paper 158* (1930), pp. 123–28.

[24] Pp. 91–92 of ftn. 19.

[25] C. P. Ross, "Correlation and Interpretation of Paleozoic Stratigraphy in South-Central Idaho," *Bull. Geol. Soc. Amer.*, Vol. XLV (1934), pp. 937–1000.

Idaho batholith) has been a positive area since about the end of the pre-Cambrian and much of the eastern, a part of the Paleozoic geosyncline that extended alongside. When the region became involved in the late Jurassic orogeny and the intrusion of the Idaho batholith (Cretaceous), it all became land and supplied erosional debris to the neighboring Laramide trough. Thus, a northeastward drainage system apparently had become established even prior to the close of the Mesozoic era. The drainage apparently was maintained through the Laramide orogeny at the close of the Cretacous period, and, as indicated by the distribution of belts of Challis volcanics (late Oligocene or early Miocene) along old northeastward-extending valleys, well into Tertiary time.

The more immediate ancestral drainage probably played an important role in post-Challis planation, and the present northeast drainage pattern may be an inheritance from late Tertiary time. The evidence afforded by headwater streams and by present wind gaps indicates that the drainage was northeast and that the Continental Divide at that time coincided closely with the watershed of the present headwater streams. Except at its very head, where there have been some subsequent adjustments to faulting, Salmon River pursued its present northeast course; but instead of making an elbow turn a short distance below the town of Salmon, it continued on in a northeasterly direction, probably along the present inverted course of Carmen Creek and on toward the present Missouri River. The Salmon was then joined by its East Fork and also by Big Lost River. The latter flowed northeast through the site of Double Springs Pass and entered the Salmon not far below the present mouth of Pahsimeroi River by way of the wind gap across the northwest corner of Lemhi Range. Antelope Creek also flowed in a northeasterly direction through what is now Wet Creek Pass and through the unnamed pass at the head of Little Lost River in the Lemhi Range, but whether it eventually entered the Salmon or continued on as an independent drainage system has not been determined. A stream no longer in existence but which may be referred to as the "Arco River" then flowed northeast through Arco Pass and Pass Creek Pass and probably on across the present Continental Divide. At that time Chamberlain Creek, Middle Fork, and Panther Creek had no local connection with Salmon River but may have joined it some miles beyond the present state line. Thus, the drainage just prior to the crustal disturbance that came at the close of the Tertiary was an integral part of a northeast drainage system that extended on into Montana and presumably to the Arctic or the Gulf of Mexico.

Disruption of this northeast drainage system probably started with the orogeny that came at the close of the Tertiary period. The general uplift apparently favored the vigorous headward erosion of a stream from the west which captured the now barbed tributaries, Chamberlain Creek and Middle Fork, then Panther Creek, and, finally, the main Salmon River. At or about the same time, rising block mountains south of the Salmon River barred the way of other northeast streams, cutting off and diverting one of its main tributaries, Big Lost River, southeast to the Snake River Plain. Antelope Creek also was forced to abandon its northeast course and to become a tributary of Big Lost River. Consequent streams developed within the structural basins, fed by streams from the bordering mountains. Warm Spring

Creek, Pahsimeroi River, and Lemhi River flowed northwest to join the Salmon. Little Lost River and Birch Creek drained southeast along the structural basins to disappear on reaching the Snake River Plain.

More or less closely associated with the mountain uplift was the Snake River downwarp.[26] While the mountains and intervening basins were being arched along an axis parallel to and about 50 miles from the margin of the Snake River Plain, the Snake River downwarp was carrying the ends of the ranges to the level of the plain and then below, where they are concealed beneath a covering of lava that partly filled the great downwarped basin. By the downwarp, the headwaters of Arco River were brought below and became a part of the underfloor of the Snake River Plain and were completely lost from view, with only Arco and Pass Creek passes left as evidence of their former existence.

Since the captures and diversions, the streams have become established in their new courses and are continuing to deepen their valleys in the mountainous areas and to aggrade their floors where they occupy the structural basins.

[26] V. R. D. Kirkham, "Snake River Downwarp," *Jour. Geol.*, Vol. XXXIX (1931), pp. 456–82.

CONCLUSIONS

Piracy and large-scale faulting have had an important role in disrupting a northeast Tertiary drainage system and in fixing the pattern of existing stream systems in the northern Rocky Mountains of east-central Idaho. Piracy has virtually reversed the direction of flow of Salmon River and turned its waters from the Missouri system to the Snake and Columbia rivers and the Pacific Ocean. Block faulting has also diverted some of the tributaries of Salmon River and other streams into structural basins extending southeast to the Snake River Plain. Downwarping involved in the origin of the Snake River Plain has caused the bordering mountains to plunge and disappear beneath the great lava-filled depression and to carry from sight the headwaters of streams that formerly drained to the northeast. As a consequence of these diversions the Continental Divide was shifted about 100 miles east of its location in late Tertiary time.

ACKNOWLEDGMENTS.—The writer wishes to take this opportunity to express publicly his gratitude to Professor O. D. von Engeln, Cornell University, for his helpful suggestions during the preparation of this report.

ACCELERATED CHANNEL EROSION IN THE CIMARRON VALLEY IN SOUTHWESTERN KANSAS[1]

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ABSTRACT

The Cimarron River formerly had a narrow channel throughout its course from the Kansas-Colorado state line to the Kansas-Oklahoma state line. Accelerated erosion began in 1914 and has increased the average width of the channel to more than 1,000 feet. Overcultivation appears to have been the principal cause of the erosion.

INTRODUCTION

During investigations of the geology and ground-water resources of Morton, Stevens, Grant, Haskell, and Seward counties, Kansas, for the state and federal geological surveys, the writer has obtained many data indicating extreme widening of the channel of the Cimarron River in southwestern Kansas. The widening was first reported by H. T. U. Smith[2] and was later described briefly by the writer.[3] In most places the accelerated channel erosion began with the disastrous flood of May 1 and 2, 1914, which destroyed many buildings and bridges as well as much of the bottom land. The principal data include testimony of the older residents of the area, photographs taken before the flood of May, 1914, and measurements of the stream channel that were made during a survey of this area in 1874. In addition, data have been obtained from files of newspapers and from histories of southwestern Kansas and of the Santa Fe trail.

The Cimarron River heads in northeastern New Mexico and flows eastward into the panhandle of Oklahoma, turns northeast across the southeastern corner of Baca County, Colorado, and enters Kansas about 5 miles north of the southwest corner of the state. Its course across southwestern Kansas is shown in Figure 1. In southern Meade County the river swings eastward for a short distance and then flows southeastward into Beaver County, Oklahoma, where it flows parallel to the Kansas-Oklahoma line until it re-enters Kansas in south-central Clark County. It then flows eastward into Commanche County, re-enters Oklahoma, and empties into the Arkansas River a short distance west of Sand Springs. The length of the Cimarron is approximately 700 miles, and the length of its course from western Morton County to southern Meade County is 144 miles.

From the Kansas-Colorado line to the first crossing of the Kansas-Oklahoma line the Cimarron River has an average gradient of 8.5 feet to the mile. The gradient is about 9 feet to the mile in

[1] Published with the permission of the director of the United States Geological Survey and of the director of the State Geological Survey of Kansas.

[2] "Geologic Studies in Southwestern Kansas," *Kan. Geol. Surv. Bull. 34* (1940), pp. 174–76; "Notes on Historic Changes in Stream Courses of Western Kansas, with a Plea for Additional Data," *Trans. Kan. Acad. Sci.*, Vol. XLIII (1940), pp. 294–300.

[3] "Geology and Ground-Water Resources of Morton County, Kansas," *Kan. Geol. Surv. Bull. 40* (1942), pp. 14–16; "Geology and Ground-Water Resources of Grant, Haskell, and Stevens Counties, Kansas," *Kan. Geol. Surv. Bull. 61* (1946), p. 36.

Morton and Stevens counties, 7 feet to the mile in Grant, Haskell, and northwestern Seward counties, and 9 feet to the mile in southeastern Seward and southwestern Meade counties. The valley bottom is less than a mile wide in most places but broadens in parts of Seward County and in southwestern Meade County. The width ranges from a little more than 1 mile in southern Grant and southwestern Haskell counties to about 4 miles in southeastern Seward and southern Meade counties. The valley probably is of about the same width in western Morton County, where the south side has a gentle slope that merges with the upland and is masked by deposits of dune sand. Deposits of high-terrace sand and gravel 2.5 miles south of the river indicate that the valley is at least 4 miles wide in that area. The valley is about 100 feet deep between the Colorado-Kansas line and southern Grant County but deepens to more than 250 feet in southeastern Seward and southern Meade counties.

The channel of the Cimarron River is wide and sandy in most parts of southwestern Kansas. The width ranges from about 140 feet at the Satanta bridge in Haskell County (Fig. 1) to more than 3,000 feet in western Morton County. In many places the sides of the channel are inconspicuous, whereas in other places they may be as high as 25 feet. The average height probably is between 5 and 10 feet.

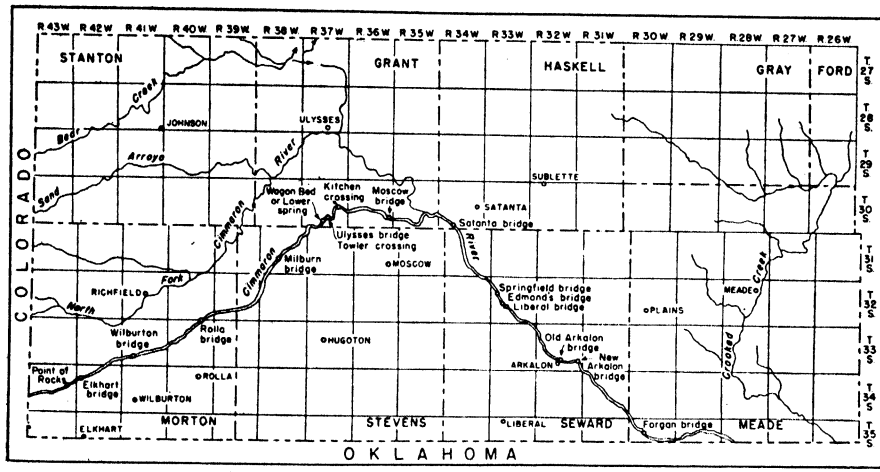


FIG. 1.—Map of southwestern Kansas showing the Cimarron River and its principal tributaries

The principal tributaries of the Cimarron are North Fork Cimarron River and Crooked Creek. North Fork Cimarron, heading in southeastern Colorado, parallels the Cimarron through Morton, Stevens, and Stanton counties into Grant County, where it swings abruptly southeastward and joins the Cimarron near the Grant-Haskell line (Fig. 1). West of this confluence, North Fork Cimarron flows at an altitude lower than that of the Cimarron; hence it probably is the older stream. The course of Crooked Creek, as far as the Oklahoma border, is seen in Figure 1.

GEOLOGY AND GEOLOGIC HISTORY

The part of southwestern Kansas through which the Cimarron River flows is underlain primarily by rocks of Tertiary and Quaternary age. The oldest rocks in the general area are Permian redbeds which crop out along Crooked Creek in Meade County, whereas the oldest rocks in the Cimarron Valley are redbeds which are believed to be Triassic and which crop out in the vicinity of Point of Rocks in western Morton County. Cretaceous deposits underlie the northern part of the area but are not visible in the Cimarron Valley. The oldest Tertiary beds are those that comprise the Laverne formation, cropping out in the Cimarron Valley in southeastern Seward and southwestern Meade counties. They are steeply dipping beds that consist of sandstone and shale in the upper part and thick deposits of sand and gravel in the lower part. They are believed to be Lower Pliocene. Deposits of the Ogallala formation, which are of Middle Pliocene age, crop out in Meade County on the east side of Crooked Creek and in Morton County in the vicinity of Point of Rocks. The Rexroad formation, which is either Upper Pliocene or Lower Pleistocene, crops out in many places in the Cimarron Valley, particularly in Seward and Meade counties, and is overlain by the Pleistocene Meade formation. Between eastern Morton County and the Liberal bridge (Fig. 1) the channel of the Cimarron River is approximately at the level of the contact of the Rexroad and Meade formations. Where the basal sand and gravel of the Meade formation lies in deep channels in the underlying Rexroad, the contact is below the level of the stream bed, and, where the sand and gravel of the Meade are thin, the contact generally is above the level of the stream bed. More recent sediments in this area include dune sand in the Cimarron Valley in Morton and Stevens counties, the alluvium of the Cimarron River, and deposits of colluvium overlying the pediment-like slopes within the valley.

There are at least two and perhaps three terraces in the Cimarron Valley in southwestern Kansas. H. T. U. Smith[4] reported terraces that are 20 and 55 feet above the flood plain in Stevens County and 20, 55, and 80 feet above the flood plain in Seward County. A prominent high-level terrace in Meade County was reported by J. C. Frye and C. W. Hibbard.[5] This terrace has been observed by the writer in many parts of Seward County and in western Morton County. Its level is only about 50 feet below the surface of the upland; hence it probably is the oldest terrace in the valley. Frye and Hibbard report that teeth of *Paraelephas columbi* (Falconer) have been taken from deposits underlying this terrace in Meade County. The remains of fossils and the fact that the terrace deposits are in channels in the Pleistocene Meade formation indicate that this terrace was formed during very late Pleistocene time and that almost all the downcutting in the Cimarron Valley was during later Pleistocene and Recent time. The presence of high-level terrace deposits near the borders of the present valley show that the flood plain during high-terrace time was approximately as wide as the valley now is and was in some places as much as four times as wide as the present flood plain.

[4] Pp. 123 and 126 of ftn. 2 (1940).

[5] "Pliocene and Pleistocene Stratigraphy and Paleontology of the Meade Basin, Southwestern Kansas," *Kan. Geol. Surv. Bull. 38*, Part 13 (1941), p. 420.

STREAM FLOW

Except during floods the flow of the Cimarron River in southwestern Kansas is governed primarily by the relation of the level of the channel to the level of the water table in the upland areas adjacent to the river (Fig. 2). At Point of Rocks the water table in adjacent upland areas is above the level of the stream bed; hence there generally is water in the channel in that vicinity. This point was known as "Middle Spring" on the old Santa Fe trail. From a short distance below Point of Rocks almost to the Rolla bridge (Fig. 1) the water table in areas adjacent to the Cimarron is very slightly below the level of the stream bed. The river in this area is a losing stream, but during periods of excessive precipitation the water table may rise sufficiently to cause a small flow in a few places.

Between the Rolla and the Milburn bridges (Fig. 1) the stream channel is slightly below the level of the water table in adjacent areas, and, hence, there generally is a small flow in that area. This flow disappears rapidly between the Milburn bridge and the Grant-Stevens line, owing to an abrupt reversal of the relationship between the stream channel and the adjacent water table.

For several miles east from the Grant-Stevens line the water table is much higher than the river channel; hence the flow in this section is perennial except perhaps during excessively long droughts. As the water moves eastward, it sinks beneath the channel because the water table becomes lower than the channel. From the Ulysses bridge to northwestern Seward County the channel lies approximately at the level of the contact of the Meade and Rexroad formations. The lower part of the Meade formation consists primarily of unconsolidated sand and gravel, whereas the upper part of the Rexroad formation is of partly consolidated to well-consolidated sand, silt, clay, and caliche. Where the upper part of the zone of saturation is in the lower part of the Meade formation, the water moves freely through the coarse sediments, and the slope of the water table is slight; where the upper part of the zone of saturation is in the upper part of the Rexroad formation, the water moves slowly through the fine-grained sediments, and the water table is steep. In addition, the gradient of the river in southeastern Grant, southwestern Haskell, and northwestern Seward counties is less than in other parts of southwest-

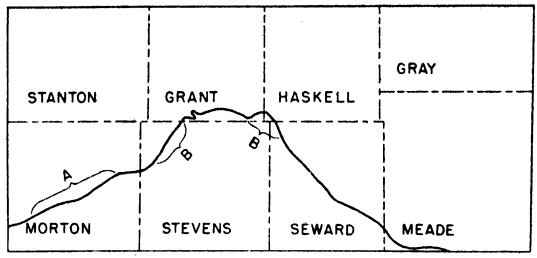


FIG. 2.—Map of southwestern Kansas showing the course of the Cimarron River. *A*, Section in which the water table and river channel are at about the same level. *B*, Sections in which the water table lies below the river channel. The water table in all other sections lies above the level of the stream.

ern Kansas, owing perhaps to the resistance to erosion of the upper part of the Rexroad formation. Because of the increased slope of the water table and the decreased gradient of the river channel, the water table is much lower than the channel in the vicinity of the Satanta bridge (Fig. 1), and, consequently, this section of the river is dry except during floods. In the vicinity of the Springfield bridge, the stream reappears, owing to an increased stream gradient and decreased slope of the water table. The slope of the water table is less in this area because the upper part of the zone of saturation is in the coarse deposits in the lower part of the Meade formation. In southeastern Seward and southern Meade counties the water table is much higher than the river channel, and the flow, therefore, is perennial.

An over-all picture of the stream flow in the Cimarron River is best shown by Figure 3, which is a graph of the stream flow based on measurements at six places in southwestern Kansas. The graph shows a net loss in stream flow between Point of Rocks and the station at Satanta bridge, with the flow generally increasing greatly from the Liberal bridge to the Forgan bridge south of Plains. The loss in stream flow from Point of Rocks to the Ulysses bridge (north of Hugoton) appears to be nearly uniform except during periods of low flow, when the loss is very slight. During periods of low flow the gradient from the surface of the water in the channel to the water table in adjacent areas is small, and little water moves laterally into adjacent areas. When the stream flow is high, the gradient is increased and more water moves into adjacent areas.

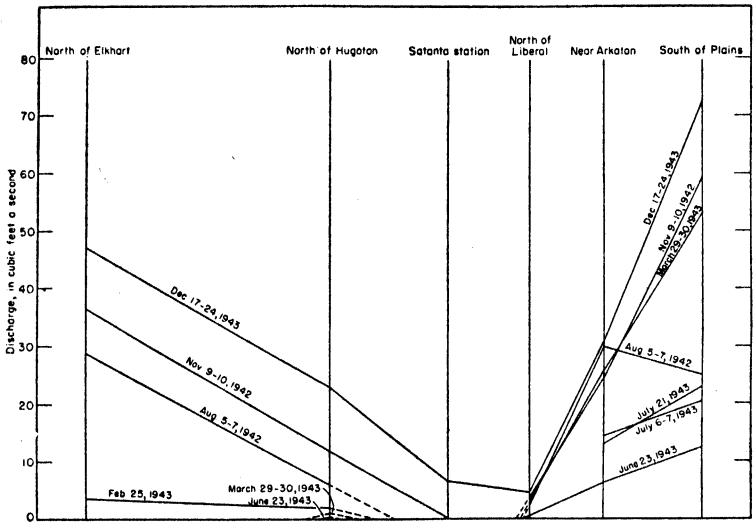


FIG. 3.—Graph of stream flow in the Cimarron River in southwestern Kansas based on measurements made by the United States Geological Survey and the states of Kansas and Oklahoma. (*Kan. Geol. Surv. Bull. 61.*)

The graph also indicates that a flow of nearly 40 cubic feet a second at Point of Rocks is required to cause a flow at

the Satanta bridge and that a flow of 45–50 cubic feet a second at Point of Rocks would leave no dry stretches in the channel in southwestern Kansas.

EARLY DESCRIPTIONS

One of the best early accounts of the Cimarron River is that of Josiah Gregg,[6] who traversed the Santa Fe trail several times before 1840. On June 19, 1831, his party reached the Cimarron at a point several miles east of Wagon Bed Spring and found the river dry. They traveled southward for one day and then northwestward until they again encountered the Cimarron (near Lower or Wagon Bed Spring), which "with its delightful green-grass glades and flowing torrent had all the aspects of an 'elysian vale,' compared with what we had seen for some time past." In the same year, when a fur-trader, Jedediah Smith, arrived at Lower Spring, "his eyes were joyfully greeted with the appearance of a small stream meandering through the valley that spread before him."[7] During a later crossing of the Cimarron, Gregg[8] says of it:

> Its water disappears in the sand and reappears again, in so many places, that some travelers have contended that it "ebbs and flows" periodically. This is doubtless owing to the fact, that the little current which may flow above the sand in the nite, or in cloudy weather, is kept dried up, in an unshaded channel, during the hot summer days . . . . In these sandy stretches of the Cimarron, ... travelers procure water by excavating basins in the channel a few feet deep, into which the water is filtrated from the saturated sand.

On June 12, 1847, when Dr. A. Wislizenus[9] reached Lower Spring, he described the Cimarron as "a small green valley, spread out like an oasis in the desert. The water is fresh and running, and the rushes grow on the banks." Ten years later, Eugene Bondel wrote that the Cimarron River above Middle Spring (at Point of Rocks) is "a narrow, running stream, which can be jumped across almost anywhere."[10] Orestes St. John,[11] who made a reconnaissance of the geology of southwestern Kansas in 1885, depicted the river thus:

> [It is a] small brook only a few yards wide, and in places during a portion of the year its waters are lost in the sandy bed. In early summer its low banks are sometimes overflowed. The above-mentioned affluents mostly afford pools the year round, and like the larger stream they are subject to overflow from the heavy local rains that occur during the summer months. In the past these pools were the resort of herds that pastured the adjacent plains, and the Cimarron Valley was, until recently, entirely occupied by stock ranches and thousands of cattle.

Erasmus Haworth[12] stated in 1897 that "the Cimarron seems to have reached base-level and to have begun meandering across its flood plain. Beautiful oxbow curves are frequent, and a sluggish nature is everywhere manifest during times of low water." Five years later (1902), W. D. Johnson[13] wrote:

> There is strong suggestion that by fan construction in the High Plains region it [the Arkansas River] has lifted itself out of a former deep valley and shifted to another course, the

[6] *Commerce of the Prairies* (New York: Henry G. Langley Co., 1844), Vol. I, pp. 74–93.

[7] *Ibid.*, p. 93. [8] *Ibid.*, Vol. II, pp. 197–204.

[9] "Memoir of a Tour to Northern Mexico in 1846 and 1847," *Sen. Misc. Doc. 26* (30th Cong., 1st sess. [1848]), p. 13.

[10] Ralph P. Bieber, *Frontier Life in the Army* (Glendale, Calif.: Arthur H. Clark Co., 1942), p. 179.

[11] "Notes on the Geology of Southwestern Kansas," *Kan. State Board of Agric., 5th Biennial Rept. for Years 1885–86* (1887), p. 133.

[12] "Underground Waters of Southwestern Kansas," *U.S. Geol. Surv. Water-Supply Paper 6* (1897), p. 22.

[13] "The High Plains and Their Utilization," *U.S. Geol. Surv., 22d Ann. Rept.*, Part 4 (1902), pp. 647 and 663–64.

deep valley thus abandoned being that now occupied by the feeble Cimarron. . . . . The valley is . . . . almost unvisited by run-off floods, yet with a perennial stream of constant volume looping intricately upon it. . . . . It does not run full length. Though in some of its live sections it is a strong stream, unvarying in volume, there are other sections in which the bed is permanently dry to depths of 20 or 30 feet. . . . . As elsewhere, under like circumstances, wild hay is cut here for winter feed. . . . . Wherever within the High Plains belt the Cimarron Valley shows a living stream it is always a meandering, looping stream of uniform width, narrow, clear, and deep.

In 1905 W. G. Russell[14] observed that, about 14 miles northwest of Hugoton, the Cimarron "is very sluggish and dirty, with little flowing water, but having in its bed many large pools. The bottoms are sand covered with straggling sage brush and a little grass, with here and there a small body of grass land from which hay is cut." The following year C. N. Gould[15] stated that, in the panhandle of Oklahoma, the Cimarron was no more than 20 feet wide. Later, H. N. Parker[16] wrote:

From the old post office at Metcalf, Okla., to Point of Rocks, Kans., a distance of 25 miles, the channel of the Cimarron is often dry, but at Point of Rocks, Kans., the water comes to the surface . . . . and the channel is usually full for a number of miles. . . . . In Kansas, from Arkalon southward, the Cimarron River usually has water in it throughout the greater part of the year. . . . . William Easton Hutchison states that the Cimarron River is a constantly running stream throughout the entire width of Morton County, where it has a valley on one side or the other of the channel . . . . on which an abundant crop of natural hay is cut. In Stevens County there is running water in Cimarron River at all seasons of the year. . . . . In Grant County the river flows constantly and has a fine fertile valley on each side of the channel that is sometimes covered by floods.

Although the above accounts contain certain minor inconsistencies, the general picture is clear.

RECENT TESTIMONY

Early residents of southwestern Kansas agree that the Cimarron River then had a narrow channel but that its width has increased greatly since the flood of 1914 and continues to increase. In Morton County, for example, E. M. Dean of Richfield stated that the river formerly was a small meandering stream and that abundant crops of natural hay were grown in the flood plain. The width of the channel at Point of Rocks was 66 feet in 1874, whereas in 1916 it was necessary to build a bridge having a length of 644 feet in order to span the channel at that place. The width of the channel in 1939 was approximately 1,400 feet (Pl. I). Several ranch buildings that were situated on the flood plain at Point of Rocks in 1937 had been engulfed by the channel by 1944. Figure 4, *A*, is a view of the Cimarron River channel at Point of Rocks in 1939.

The widening of the channel has been particularly great in some parts of northwestern Stevens County. The history of the channel in sec. 23, T. 32 S., R. 39 W. is typical of those areas where widening has been very great. The width of the channel at the east line of section 23 in 1874 was 36 feet (probably measured in the direction of the line of the survey rather than normal to the direction of stream flow). In 1930, when the Panhandle Eastern Pipeline Company constructed a pipeline across the channel in section 23, the channel had widened to

[14] "Investigations in Kansas," *House Doc. 44* (58th Cong., 2d sess. [1904]), p. 327.

[15] "Geology and Ground-Water Resources of Oklahoma," *U.S. Geol. Surv., Water-Supply Paper 148* (1905), p. 90.

[16] "Quality of the Water Supplies of Kansas," *U.S. Geol. Surv. Water-Supply Paper 273* (1911), pp. 306–7.

PLATE I

Aerial view of Cimarron River in the vicinity of Point of Rocks (lower left corner at *N*) and Elkhart bridge (upper right center). Note narrow patches of alluvium (darker areas) along north side of valley floor that have not yet been destroyed by the widening channel. The channel has almost engulfed the small island of alluvium downstream from Elkhart bridge. Arrow is about a mile long. (Photograph by U.S. Department of Agriculture.)

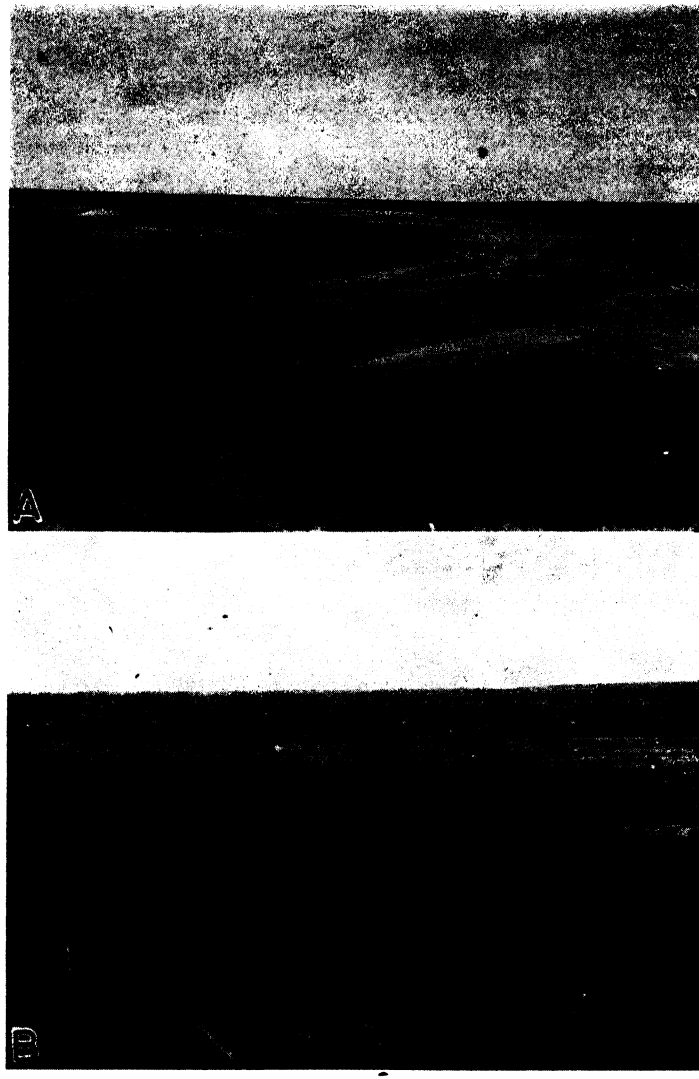


FIG. 4.—*A*, Cimarron River at Point of Rocks in 1939. (Photograph by H. A. Waite.) *B*, Flood plain of Cimarron River in 1899. Note the narrow, steep-walled channel in the foreground. (Photograph by W. D. Johnson.)

1,200 feet.[17] During the next year the same company drilled a gas well in section 23, and, although the well was located on the flood plain about 300 feet north of the river channel, the widening of the channel was so great that by July, 1942, the well was in the stream bed at a point about 410 feet south of the north bank of the channel (Pl. II). The left wall of the channel, therefore, has been shifted 710 feet in about 11 years.

The widening of the channel has not been so great in Grant and Haskell counties as in Morton and Stevens counties, owing to the narrowness of the flood plain in this section and to the decrease in stream gradient. Mrs. Oscar McClure of Hugoton stated that her father built one of the first bridges across the Cimarron in southern Grant County. The bridge, located southwest of Wagon Bed Spring, was 32 feet long, whereas the channel was about 500 feet wide at that point in 1943. G. E. Gordon of southern Grant County used a 20-foot plank for a foot bridge across the channel of the Cimarron before it increased to its present width of more than 400 feet.

R. R. Wilson[18] of Garden City wrote the following about the Cimarron River:

> I am satisfied that the channel was not over 30 feet wide there [at the Ulysses bridge] in 1886 and was mud bottom. My father built a bridge across the river just a few rods above where the Ulysses bridge is now located in the spring of 1886 and I remember this was a very short bridge. . . . . There was always water in the river from Wagon Bed Spring down stream for about 10 miles, until along about 1897 or 1898 I camped on my homestead about 6 miles below Wagon Bed Spring and that summer was very dry and the water would all disappear about noon and through the afternoon, then every morning there would be a stream of water running. . . . . I am quite sure the channel was always dry at the Satanta bridge . . . . , also the channel was nearly always dry above Wagon Bed Spring for some distance.

Views of the Cimarron River at Towler crossing (site of Ulysses bridge) in 1903 and in 1943 are shown in Plate III.

Judge William Easton Hutchison,[19] of Garden City, writes:

> The bed of the stream there, the Kitchen crossing, was in my best recollection about 10 feet wide, surely not more than 12 feet. There was always water covering the bed of the channel, I would say 6 or more inches deep. . . . . I frequently drove from Ulysses to . . . . Liberal in Seward County, but I kept north of the river and didn't cross it until I reached the Edmonds crossing north of Liberal about 6 or 8 miles. There seemed to be a little more water there than at the Kitchen crossing in Grant County—maybe 10 or 12 inches deep but not much wider than at the Kitchen crossing. I also crossed this stream at Arkalon occasionally. That was very much like the Edmonds crossing.

J. W. Dappert,[20] of Taylorville, Illinois, county surveyor of Grant County in 1887 and 1888, writes as follows:

> I had been up and down the Cimarron River in Commanche County, Kansas and . . . . in the Cherokee strip [northern Oklahoma], where the river was then [1886] from 200 to 600 feet in width with a very sandy bed; but up here [Wagon Bed Spring] we found a little brook with water holes at places 30 or 40 feet wide and 100 to 200 feet long; but between the water holes a narrow channel ran along such as I could and did easily step across, with a small stream of pure sparkling water flowing along in the channel. . . . . Along the Cimarron River grew a sort of slough grass, heavy in foliage, rank and green. . . . . Also flags or cat-tails grew out of the lowest parts of the marsh area of some 20 acres which was a part of the Dick Joyce homestead [NE ¼ sec. 33, T. 30 S., R. 37 W.].
>
> Again, on this trip [May 14, 1916], I took special notice of the great changes in the condition and format of the Cimarron River. I had known it when I could easily step across its channel—only 3 or 4 feet wide and 4 or 5 feet

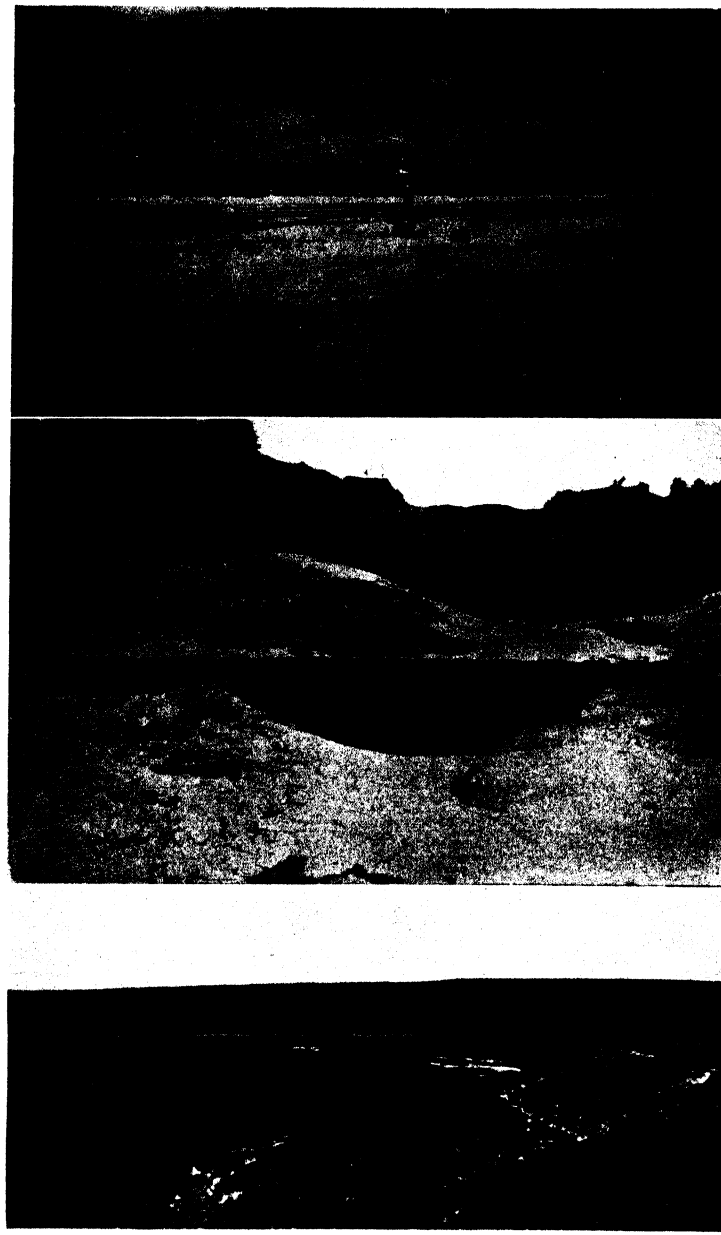
[17] Data from Samuel Neithercoate of the Panhandle Eastern Pipeline Co.

[18] Letter dated April 1, 1943.

[19] Letter dated April 2, 1943.

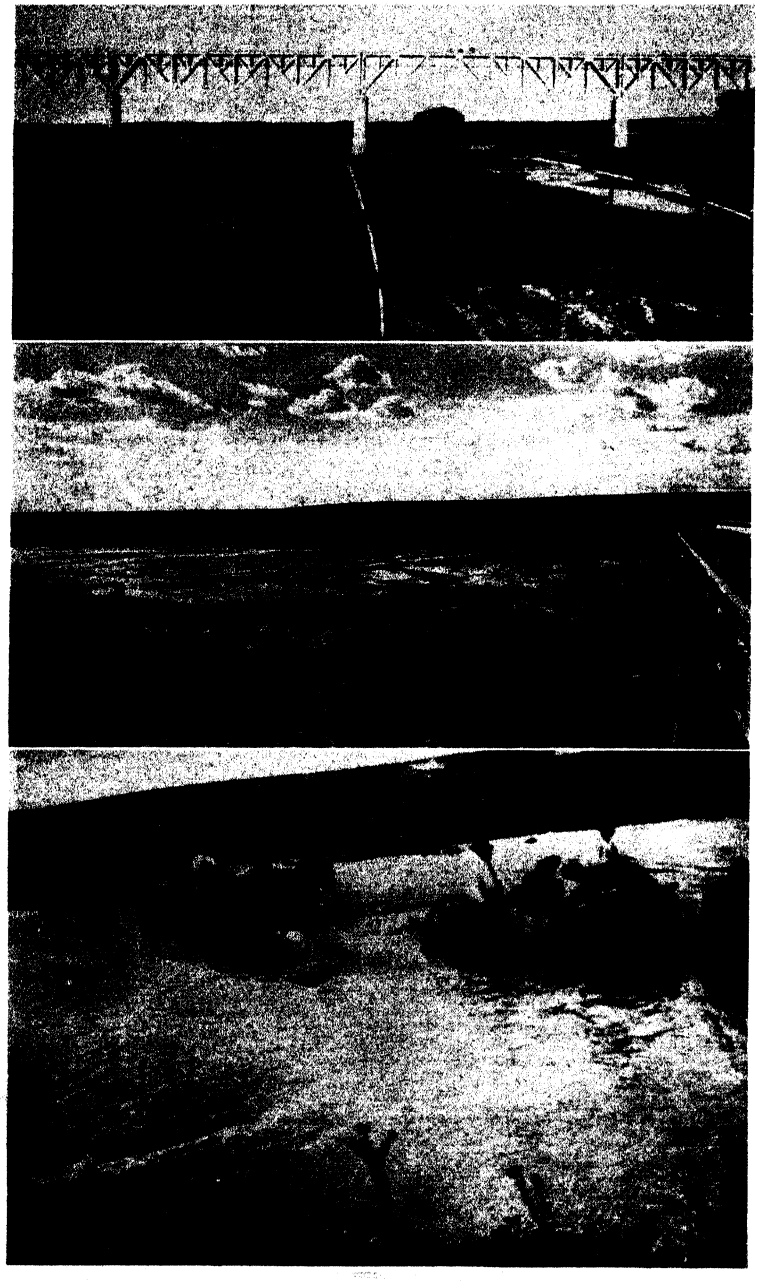
[20] Letter dated January 1, 1944.

PLATE II



A, Gas well in channel of the Cimarron River in sec. 23, T. 32 S., R. 39 W., Stevens County. *B*, North wall of channel of Cimarron River in sec. 23, T. 32 S., R. 39 W., Stevens County. Widening of channel has exposed dune sand overlying alluvium. *C*, Hay field on flood plain of the Cimarron in sec. 25, T. 34 S., R. 31 W., Seward County. This is one of the few places where the hay meadows have not been completely destroyed. Tilted beds of the Laverne formation in the foreground.

PLATE III



A, Cimarron River at New Arkalon bridge in 1943. White lines indicate width of channel in 1939. *B*, Towler crossing in 1943. Ranch buildings are on the brink of the channel wall. *C*, Towler crossing during the flood of 1903. Note gentle slope from ranch buildings to stream. (Photograph by D. D. Crotts.)

deep— . . . . and now, ... by actual measured [by pacing] distance, the width of the sandy bed was 450 feet [near Wagon Bed Spring] . . . . with a few trickles of water visible and flowing feebly along in the sandy bed. . . . . At Wagon Bed Spring the river had receded some hundreds of feet farther to the east, thus shortening the quite abrupt bend. . . . . I made a sketch plat of the stream showing its location and width; and later superimposing my earlier map [1886] upon my later map I noted the difference in the character of the stream. . . . . At first view there was no sand visible at all, while at its later view there was little else visible but sand.

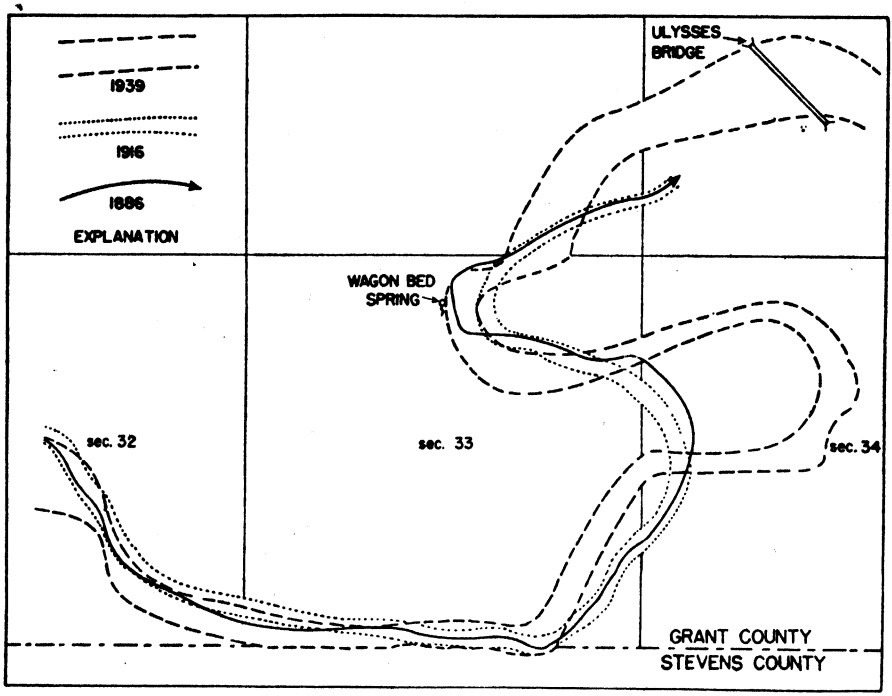


FIG. 5.—Map of a part of T. 30 S., R. 37 W., showing the Cimarron River as sketched by J. W. Dappert in 1886 and 1916 and as taken from aerial photographs made in 1939.

Figure 5 is a map of the vicinity of Wagon Bed Spring showing the position and width of the Cimarron in 1886, 1916, and 1939.

Few data concerning the width of the channel in Haskell County before 1914 are available. The widening of the channel has been less in Haskell County than in any other part of southwestern Kansas, owing perhaps to the position of the water table and the relatively low gradient of the stream. The average width of the channel in that area in 1874 was 32 feet (based on measurements at five places), whereas the width ot the Satanta bridge in 1937 was 64 feet (measured by H. T. U. Smith) and in 1942 was 140 feet.

The destruction of valuable grazing and meadow land by the widening channel of the Cimarron probably has been greatest in Seward and Meade counties, although the widening has been less there than in Morton and Stevens counties. In

Meade and Seward counties the wider flood plain afforded a much greater area of hay meadows prior to the widening. In the vicinity of the Liberal bridge the channel has widened from only a few feet before 1890 (according to Bernard Lemert) to 790 feet in 1939.

T. V. Harvey of Seward County supplied photographs that show the radical changes in the channel between Liberal bridge and Arkalon (Pl. IV). In 1939 the width of the channel at the site of photograph *A* was 1,760 feet and at the site of photographs *B* and *C* was 1,100 feet.

According to Lee Larrabee of Liberal, many acres of bottom land have been destroyed in the vicinity of Arkalon. Photographs taken by him during the flood of May, 1914, show many stacks of hay in the valley bottom (Pl. V) indicating a relatively narrow channel and a broad flat flood plain. Plate V, *C*, shows the broad sandy channel in that vicinity in 1943.

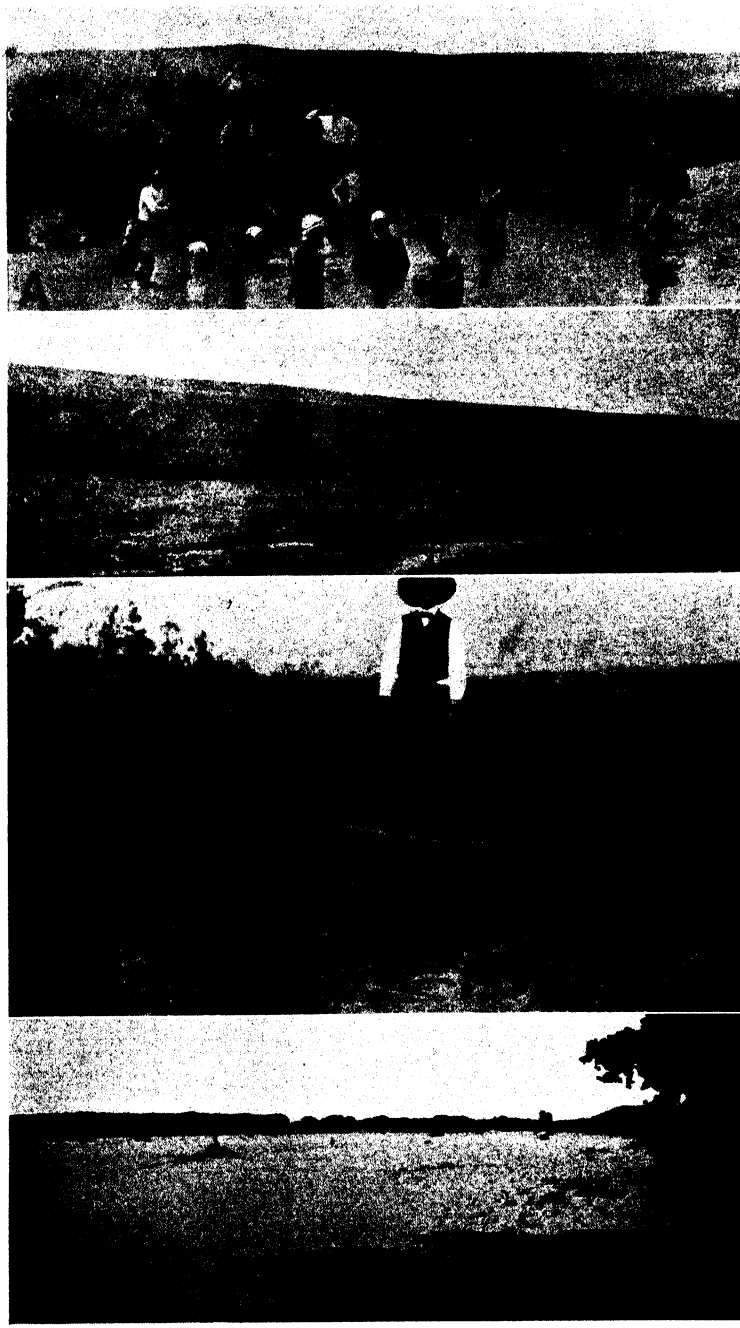
Henry Cochran, foreman of the XI Ranch in Meade County, stated that he had cut hay where there is now only a broad sandy channel. Mr. William Sourbeer reported to John C. Frye of the Kansas Geological Survey that he formerly transported his well-drilling machine across the channel of the Cimarron on two 30-foot planks. The width of the channel at the narrowest point in Meade County in 1939 was about 500 feet.

SURVEY OF 1874

The land network of most of southwestern Kansas was surveyed in 1874 during which time the width of the channel of the Cimarron River was measured at almost every section line that crossed the stream. The measurements apparently were made on the section lines rather than normal to the stream flow. This is indicated by the measurement made at the east line of sec. 34, T. 30 S. R. 37 W., Grant County, where travelers of the Santa Fe trail and early settlers in Grant County reported a very narrow channel. The surveyors reported a width of 132 feet, indicating that their measurement was not made normal to the stream which at that point flows almost parallel to the north-south section line. The measurements of 1874 generally were listed to the nearest 5 or 10 links. The greatest width reported by them was 304 feet in southern Meade County, the least was 10 feet in T. 32 S., R. 33 W., Seward County, and the average width was about 51 feet. The channel was widest in the few miles before the river entered Oklahoma. The width in this section (the last two townships through which the river flows in Meade County) averaged 154 feet, whereas above this section it averaged 45 feet. In Table 1 are listed the widths of the channel in 1874 and in 1939. Measurements listed in italics were taken from contact prints of aerial photographs and probably are accurate to within 10 or 15 feet, whereas all other measurements in 1939 were determined from aerial mosaics and probably are accurate only to within 100 feet. The greatest average width in 1939 was about 2,000 feet in Morton County, the smallest average width was about 480 feet in Haskell County, and the average width from the Colorado line to the Oklahoma line was about 1,160 feet (based on 156 measurements).

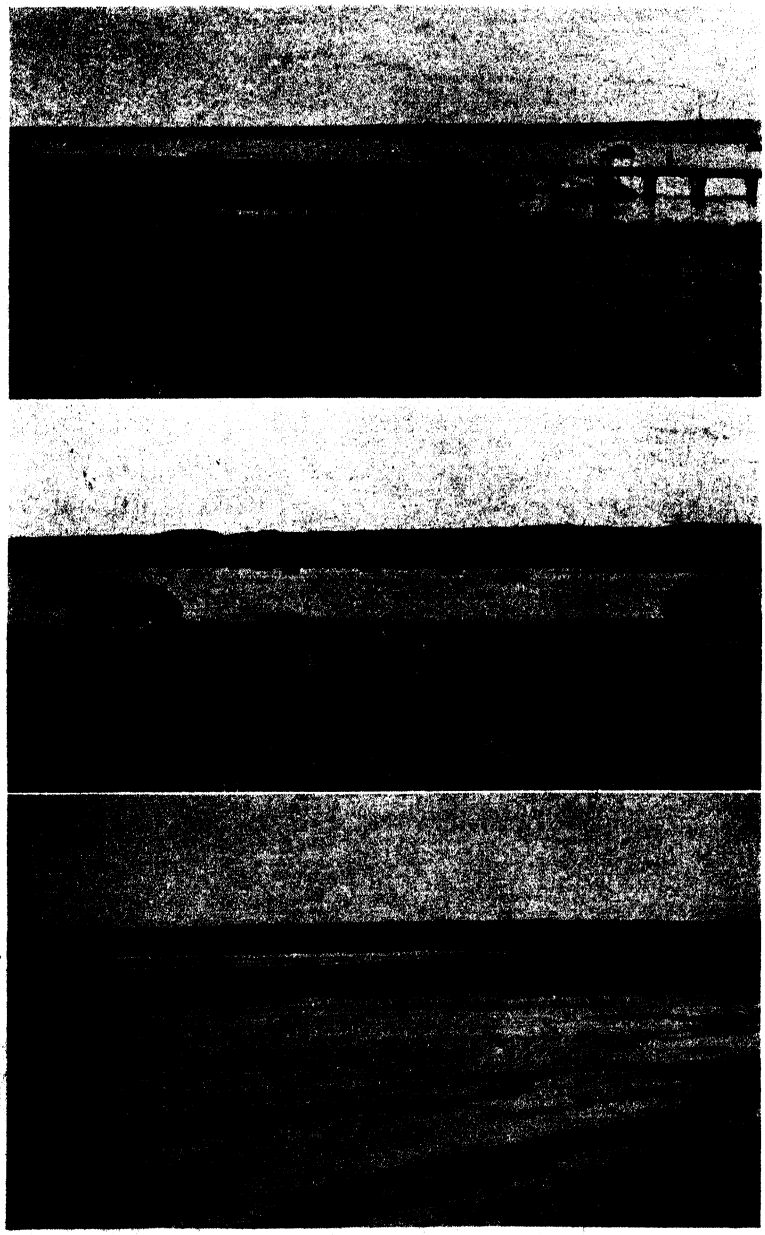
ACCELERATED EROSION IN TRIBUTARY STREAMS

There has been comparatively little channel erosion in the major tributaries of the Cimarron River owing primarily to the lower stream gradients. According to



A, Channel of Cimarron in sec. 20, T. 33 S., R. 32 W. in 1897. (Photograph by Lee Larrabee.) *B*, View southwestward across Cimarron Valley in NW¼ sec. 18, T. 33 S., R. 32 W. in 1907. Heavy foliage in center marks position of channel. (Photograph by T. V. Harvey.) *C*, Close-up view of channel at a point near the left end of *B*. (Photograph by T. V. Harvey.) *D*, Cimarron channel in 1943. View northward from point about 100 yards downstream from left end of *B*.

PLATE V



A, Cimarron Valley at Old Arkalon bridge during flood of May 2, 1914. (Photograph by Lee Larrabee.)
B, Cimarron Valley about a mile below Arkalon during flood of May 2, 1914. (Photograph by Lee Larrabee.)
C, Cimarron Valley about a mile below Arkalon in 1943.

TABLE 1

WIDTHS OF THE CHANNEL OF THE CIMARRON RIVER IN 1874 AND 1939

Location	Width in 1874 (Feet)	Width in 1939 (Feet)
MORTON COUNTY:		
T. 34 S., R. 43 W.		
West line sec. 19	53	2,900
West line sec. 20	53	3,000
West line sec. 21	46	3,200
West line sec. 22	40	2,800
North line sec. 22	33	2,100
West line sec. 14	79	1,800
East line sec. 14	66	2,300
North line sec. 13	.....	2,400
T. 34 S., R. 42 W.		
West line sec. 7	139	1,300
East line sec. 7	66	*2,150*
North line sec. 8	.....	*2,380*
East line sec. 5	59	*1,400*
East line sec. 4	46	2,600
North line sec. 3	73	1,800
T. 33 S., R. 42 W.		
West line sec. 35	46	3,200
North line sec. 35	46	800
East line sec. 26	46	800
South line sec. 25	66	1,100
East line sec. 36	.....	800
T. 33 S., R. 41 W.		
South line sec. 30	66	1,800
West line sec. 29	66	2,200
West line sec. 28	66	2,600
East line sec. 28	46	1,100
South line sec. 22	66	1,500
West line sec. 23	66	2,000
West line sec. 24	66	2,400
NE corner sec. 24	46	2,500
T. 33 S., R. 40 W.		
East line sec. 18	.....	1,700
East line sec. 18	.....	1,600
East line sec. 18	33	300
South line sec. 8	33	2,100
West line sec. 9	66	2,900
SW corner sec. 3	56	2,400
NE corner sec. 3	43	2,600
T. 32 S., R. 40 W.		
NE corner sec. 35	33	1,100
T. 32 S., R. 39 W.		
West line sec. 30	46	2,600
East line sec. 30	53	1,300
East line sec. 29	53	1,400
Av. in Morton County	56	2,000
STEVENS COUNTY:		
T. 32 S., R. 39 W.		
West line sec. 27	53	2,200
West line sec. 26	46	2,500
North line sec. 26	46	2,100
NE corner sec. 23	36	2,200
North line sec. 13	33	1,800
North line sec. 12	40	700
East line sec. 1	.....	1,900
STEVENS COUNTY—*Continued*		
T. 31 S., R. 38 W.		
South line sec. 31	.....	1,300
SW corner sec. 29	46	1,300
SW corner sec. 21	40	1,700
North line sec. 21	46	1,000
West line sec. 15	40	1,100
NE corner sec. 15	46	800
North line sec. 11	.....	1,000
East line sec. 2	.....	800
North line sec. 1	.....	600
Av. in Stevens County	43	1,400
GRANT COUNTY:		
T. 30 S., R. 37 W.		
West line sec. 32	26	600
West line sec. 33	40	500
West line sec. 34	40	600
South line sec. 28	40	500
West line sec. 27	50	800
North line sec. 34	66	800
West line sec. 35	132	300
South line sec. 26	99	400
South line sec. 23	66	500
West line sec. 24	66	300
North line sec. 25	66	700
NE corner sec. 25	.....	200
T. 30 S., R. 36 W.		
North line sec. 19	.....	500
West line sec. 17	.....	300
North line sec. 20	.....	400
West line sec. 21	66	200
West line sec. 22	66	300
West line sec. 23	66	500
South line sec. 23	.....	300
East line sec. 26	.....	200
East line sec. 25	.....	1,000
T. 30 S., R. 35 W.		
West line sec. 29	33	1,800
West line sec. 28	50	1,100
North line sec. 33	50	1,300
West line sec 34	26	1,300
South line sec. 27	26	1,300
South line sec. 22	33	800
SW corner sec. 23	17	1,100
North line sec. 26	33	700
West line sec. 25	43	500
North line sec. 25	.....	300
East line sec. 24	50	600
Av. in Grant County	52	600
HASKELL COUNTY:		
T. 30 S., R. 34 W.		
North line sec. 30	33	800
East line sec. 30	33	500
South line sec. 29	33	500

TABLE 1—*Continued*

Location	Width in 1874 (Feet)	Width in 1939 (Feet)
HASKELL COUNTY—*Continued:*		
East line sec. 32	26	300
South line sec. 33	33	300
Av. in Haskell County	32	500
SEWARD COUNTY:		
T. 31 S., R. 34 W.		
North line sec. 8	66	300
North line sec. 17	66	300
West line sec. 16	132	500
North line sec. 21	33	600
North line sec. 28	20	1,400
West line sec. 27	66	1,000
North line sec. 34	132	1,000
West line sec. 35	43	1,500
West line sec. 36	33	1,600
South line sec. 36	13	1,000
T. 32 S., R. 33 W.		
West line sec. 6	16	1,100
North line sec. 7	15	1,200
North line sec. 18	16	300
West line sec. 17	16	1,000
North line sec. 20	18	1,000
West line sec. 21	18	600
North line sec. 28	16	1,100
West line sec. 27	16	900
North line sec. 34	13	500
West line sec. 35	10	1,000
West line sec. 36	13	1,300
South line sec. 36	20	500
T. 33 S., R. 32 W.		
West line sec. 6	49	500
North line sec. 7	13	700
North line sec. 18	33	1,800
South line sec. 18	13	440
East line sec. 19	.....	260
East line sec. 20	20	1,410
South line sec. 21	26	370
West line sec. 27	26	790
West line sec. 26	20	1,100
West line sec. 25	33	350
North line sec. 36	33	1,320
East line sec. 36	33	1,180
T. 33 S., R. 31 W.		
South line sec. 31	33	880
SEWARD COUNTY—*Continued:*		
T. 34 S., R. 31 W.		
West line sec. 5	20	1,320
North line sec. 8	.....	910
West line sec. 9	20	270
North line sec. 16	20	1,140
West line sec. 15	26	260
North line sec. 22	26	1,230
SE corner sec. 15	33	970
SE corner sec. 23	30	620
East line sec. 25	33	570
Av. in Seward County	32	900
MEADE COUNTY:		
T. 34 S., R 30 W.		
South line sec. 30	33	1,050
South line sec. 31	79	570
T. 35 S., R. 30 W.		
East line sec. 6	40	1,320
South line sec. 5	43	1,410
East line sec. 8	.....	1,400
South line sec. 9	53	530
East line sec. 16	33	1,250
East line sec. 15	33	800
East line sec. 14	66	1,000
East line sec. 13	132	1,000
T. 35 S., R. 29 W.		
North line sec. 18	.....	1,700
East line sec. 7	.....	1,100
South line sec. 8	.....	1,400
North line sec. 17	132	1,300
East line sec. 8	100	1,100
East line sec. 9	93	1,600
East line sec. 10	145	1,000
East line sec. 11	304	1,300
East line sec. 12	.....	1,300
East line sec. 13	132	800
T. 35 S., R. 28 W.		
South line sec. 18	172	800
Av. in Meade County	99	1,100
Av. through all counties	51	1,160

J. C. Frye,[21] the gradient of Crooked Creek is about 5 feet to the mile upstream from Fowler, about 3 feet to the mile between Fowler and Meade, and about 8 feet to the mile downstream from Meade. Similarly, North Fork Cimarron River has a lower gradient than the Cimarron River; through most of Grant County it is between 5 and 7 feet to the mile. Near its confluence with the Cimarron River, however, its gradient is much steeper, and some channel erosion has occurred there.

Channel erosion has been very extensive in the small tributary streams with-

[21] "Geology and Ground-Water Resources of Meade County, Kansas," *Kan. Geol. Surv. Bull. 45* (1942), p. 14.

in the Cimarron Valley. These streams are most abundant in the wider section of the Cimarron Valley between the Haskell-Seward county line and the Kansas-Oklahoma line. They head in the relatively steep bluffs near the edge of the upland and flow down the broad pediment-like slopes which are covered with deposits of colluvium. These short tributaries were once narrow and shallow, but most of them have been enlarged greatly since 1914. One short tributary (sec. 24, T. 34 S., R. 31 W., Seward County) was formerly crossed by a wagon road and later by an ungraded auto road. In 1943 its channel was approximately 6 feet deep and 50 feet wide.

The channels of these tributary streams appear to have been enlarged headward. One short tributary about 3 miles below the Liberal bridge has an enlarged channel only in the lower half of its course. An auto road that formerly crossed the stream along a section line now crosses at a point about ¼ mile upstream, just above the head of the large gully. Along this ¼-mile section of the stream are the remnants of several roads each of which began at the old section-line road and crossed the tributary at successively greater distances upstream.

Reports of the early travelers and settlers indicate that the channel of the Cimarron was narrow and shallow. Its walls now are as much as 30 feet high. This, however, is no certain indication that the channel has been lowered. Before accelerated erosion began, the deposits of colluvium covered the pediment-like valley slopes, as well as much of the alluvium, and extended nearly to the Cimarron River. As the channel was widened by cutting laterally into the colluvium, the channel walls could become progressively higher even without downcutting.

Downcutting in the channel of the Cimarron, however, is indicated by the rapid development of tributary gullies and by the upstream migration of the locus of reappearing stream flow below the Satanta bridge. The rapid headward erosion of the short tributary gullies indicates an increased gradient caused by a lowering of the main channel. The reappearance of stream flow below the Satanta bridge was about a mile farther upstream in 1943 than it was prior to 1914.[22] This indicates that the channel is now much lower with respect to the water table than before 1914. Some difference might be caused by a rise in water table, but a relative change in level amounting to about 10 feet cannot be explained solely by a rising water table. Ground-water studies in this area indicate very little change in the water table since 1890 except for seasonal fluctuations. It appears, therefore, that the upstream movement of the point of reappearance of stream flow was caused in part by a lowering of the channel.

CAUSES OF ACCELERATED EROSION

Accelerated channel erosion is common in the arid Southwest, where several detailed studies have been made by C. W. Thornthwaite, C. F. S. Scharpe, and E. F. Dosch[23] and by Kirk Bryan.[24]

[22] Lee Larrabee, personal communication.

[23] "Climate and Accelerated Erosion in the Arid and Semi-arid Southwest, with Special Reference to the Polacca Wash Drainage Basin, Arizona," *U S. Dept. Agric. Tech. Bull. 808* (1942).

[24] Date of Channel Trenching (Arroyo Cutting) in the Arid Southwest," *Science*, Vol. LXII (new ser., 1925), pp. 338–44; "The Papago Country, Arizona: A Geographic, Geologic, and Hydrologic Reconnaissance with a Guide to Desert Watering Places," *U.S. Geol. Surv. Water-Supply Paper 499* (1925); "Channel Erosion of the Rio Salado, Socorro County, New Mexico," *U.S. Geol. Surv. Bull. 790* (1927), pp. 17–19; and "Historic Evidence on Changes in the Channel of Rio Puerco, a Tribu-

These studies show that the greatest incidence of channel erosion was during the period from 1880 to 1900. Enlargement of the channel of the Cimarron River began with the flood of May 1 and 2, 1914. Big Sandy Creek (another tributary of the Arkansas River) has eroded similarly since about the same time, but to a lesser extent.

The cause, or causes, of accelerated valley erosion in the arid and semiarid Southwest are not known, and the theories are conflicting. The rejuvenation of streams could be a possible cause. Local rejuvenation, such as that resulting from stream capture, would not account for the widespread accelerated erosion. A regional uplift would cause rejuvenation of streams flowing down the dip of the beds but would decrease the gradient of streams flowing in the opposite direction.[25]

Proponents of the theory that a change in climate is the cause of accelerated erosion believe that a decrease in precipitation causes the vegetation to be less resistant to run-off, and hence, increases erosion. Others believe that a decrease in precipitation leads to a depletion of vegetation and that the resulting increase of sediment overloads the streams, causing them to aggrade rather than degrade their channels. Thornthwaite, Sharpe, and Dosch point out:

> Decrease of precipitation in an arid or semiarid land reduces the vigor of plant growth. The less hardy specimens die, the survivors are reduced in size or strength, and there is a general shift toward a more xerophytic vegetation. With decreasing precipitation there is also less water to flow off the lands. Three results are possible: (1) If the effect of vegetal depletion is less than that of reduction in precipitation, run-off will be less destructive and the hypothetical change of climate will bring less erosion than formerly; (2) vegetal depletion may balance decrease in precipitation and erosion conditions may remain the same as before; (3) depletion of vegetation may be more effective than the decrease in precipitation and, particularly if individual storm intensity remains the same, run-off and erosion will be accelerated.[26]

Bryan and others believe that a climatic change has been the principal cause of accelereted erosion in the arid Southwest. As evidence, Bryan states:

> On Chaco River . . . . an arroyo has been cut since 1860. Previously, however, an arroyo similar to the modern arroyo was formed in late prehistoric time. This channel was filled with alluvium and the floor of the valley once more became a true flood plain. Obviously the erosion of this prehistoric arroyo could not have been brought about by overgrazing. On Rio Puerco a similar ancient arroyo, completely filled by renewed sedimentation, has been discovered. . . . . Thus on both of these streams cyclic changes from sedimentation to erosion and back to sedimentation took place before the introduction of livestock and the formation of the existing arroyos. It appears inherently most probable that these cyclic changes have a common, and doubtless a climatic, cause. The introduction of livestock and the ensuing overgrazing should be regarded as a mere trigger pull which timed a change about to take place.[27]

Others have observed gullying in places where cattle had never been grazed.

Thornthwaite, Sharpe, and Dosch[28] concluded that the prehistoric channel erosion described by Bryan probably was caused by extreme climatic fluctuations such as could be expected only once in 100 years or perhaps only once in 500 years.

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tary of the Rio Grande in New Mexico," *Jour. Geol.*, Vol. XXXVI (1928), pp. 265–82.

[25] P. 107 of ftn. 23.

[26] P. 109 of ftn. 23.

[27] P. 281 of ftn. 24 (1928).

[28] P. 46 of ftn. 23.

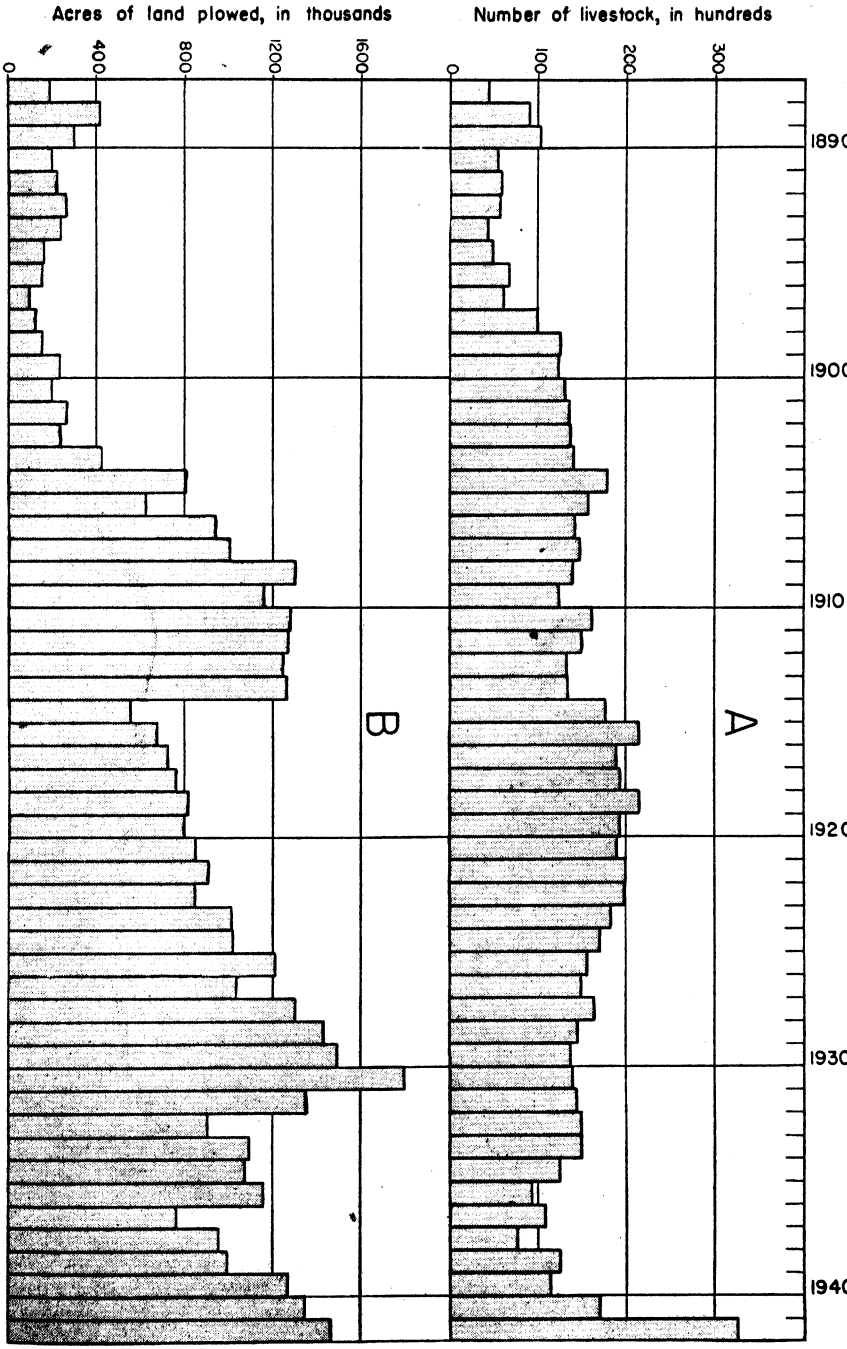

FIG. 6.—*A*, Livestock in southwestern Kansas between 1888 and 1942. *B*, Acres of land plowed in southwestern Kansas between 1888 and 1942.

Overgrazing commonly has been thought to be the principal cause of accelerated erosion in the arid regions of southwestern United States. Accelerated erosion began in many streams in that area soon after the introduction of large herds of cattle which overgrazed the pastures. In the Cimarron Valley, however, grazing appears to have had little or no effect upon erosion. Figure 6, *A*, shows the number of livestock in that part of southwestern Kansas which is drained by the Cimarron River. (Similar data for the rest of the Cimarron drainage basin were not available.) There was no sharp increase in livestock population in or near 1914; hence, it can be concluded that overgrazing was neither the cause nor the "trigger pull" that brought about erosion of the channel of the Cimarron River.

The overcultivation of land is not considered a cause of accelerated erosion in the arid Southwest, which has always been primarily a grazing country. It appears, however, that overcultivation may be the principal cause of erosion in the Cimarron Valley. Figure 6, *B*, shows the acres of land under cultivation in southwestern Kansas from 1888 to 1942. The acreage of plowed land averaged about 250,000 in that area from 1888 to 1903. From 1903 through 1908 the acreage increased rapidly and from 1909 through 1914 averaged about 1,250,000. After 1914 much of the land lay idle without care and without a protective vegetal covering. It was not until 1928 that the acreage of plowed land exceeded that of the period from 1909 to 1914. It is, therefore, believed that the rapid growth of agriculture in this area from 1909 to 1914 and the subsequent abandonment of thousands of farms was the *immediate* cause of accelerated erosion in the Cimarron Valley.

SUMMARY AND CONCLUSIONS

The evidence here presented proves definitely that the Cimarron River formerly had a narrow channel (meandering in some places) in most of its course between the Kansas-Colorado line and the Kansas-Oklahoma line. Below south-central Meade County the channel was much wider as is indicated by the report of J. W. Dappert and by photographs in the library of the Kansas Historical Society which show that the Cimarron had a moderately wide sandy channel in Clark County prior to 1900. Throughout the river's course in southwestern Kansas there were many areas of flat bottom land on which natural hay was grown, but today there are only a few remnants of these hay fields in southeastern Seward and southwestern Meade counties. The reports on the extent of early floods are conflicting. A few reported that the valley was almost never visited by floods, whereas others said that they were common. Early photographs (Pls. III and V) indicate that floods spread over much of the flood plain, but recent floods have been confined primarily to the channel, inasmuch as the channel now occupies most of the valley floor.

The date of the beginning of extensive widening of the channel was May, 1914, when a flood (Pl. V) destroyed much bottom land throughout the course in southwestern Kansas. The widening has continued since that time and was particularly great during floods in 1941 and 1942. By 1943 all the wooden bridges in this area had been destroyed, and the channel had widened beyond one end of many of the newer steel and concrete bridges. A wooden approach to the Arkalon highway bridge had to be built after a flood in 1941, when the channel

widened beyond the end of the bridge. The damage to bridges has been so great that the Chicago, Rock Island and Pacific Railway Company constructed, at a cost of more than a million dollars, a steel span that extends almost entirely across the flood plain and which probably will not be affected by widening of the channel (Pl. III, *A*). The Elkhart, Rolla, Satanta, and Arkalon bridges are the only highway bridges that have not been destroyed by floods.

The data indicate that the channel of Cimarron River has been lowered but the extent of lowering is not known.

Accelerated erosion in the Cimarron Valley is believed to be the result of overcultivation of land and the subsequent abandonment of thousands of farms, leaving large areas with lessened resistance to run-off.

ACKNOWLEDGMENTS.—The manuscript was read critically by S. W. Lohman and H. T. U. Smith of the United States Geological Survey; by John C. Frye and R. C. Moore of the State Geological Survey of Kansas; and by Claude W. Hibbard of the University of Kansas. I wish also to acknowledge the helpful co-operation of county officials and the many pioneer residents of southwestern Kansas who supplied photographs and many other pertinent data.

THE CATOCTIN FORMATION IN CENTRAL VIRGINIA[1]

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ABSTRACT

The Catoctin formation in central Virginia consists of an undetermined thickness of hydrothermally altered andesite classed as propylite. It is underlain by a variable thickness of both metamorphosed and unmetamorphosed conglomerate, sandstone, tuff, and andesite to which the name "Oronoco formation" is assigned. The Catoctin in northern Virginia is overlain by the Lower Cambrian Loudoun formation. In central Virginia, the Unicoi formation, which is at least partly equivalent to the Loudoun formation, crops out along the western flanks of the Blue Ridge. It is not exposed in the Catoctin outcrop belt because of erosion, faulting, or nondeposition. The Unicoi contains conglomerates, sandstones, tuffs, and an andesite flow lithologically like these rocks in the Oronoco and Catoctin formations. Thus, the andesitic effusives and tuffs in the Oronoco, Catoctin, and Unicoi formations are thought to be genetically related. Furthermore, the Catoctin and Oronoco rest upon a surface of major unconformity. This unconformity underlies the Unicoi. The Catoctin and Oronoco are, consequently, thought to be Cambrian in age. In the Western part of the Piedmont province, the Catoctin overlies the Lynchburg formation. The Lynchburg may therefore be partly equivalent to the Oronoco and thus of Cambrian age.

INTRODUCTION

W. B. Rogers[2] was probably the first geologist to recognize (1836) the igneous origin of the rocks later differentiated as the Catoctin formation. He thought that they were intrusive. G. H. Williams[3] first classified them as volcanic. H. R. Geiger and Arthur Keith[4] first applied the name "Catoctin schist" to these rocks in the legend of a sketch map accompanying a paper published in 1891. The formation was first described in detail by Arthur Keith.[5]

The age of the Catoctin has been in dispute since Keith[6] described it as the oldest formation exposed in the pre-Cambrian complex in the Blue Ridge province in northern Virginia. A. S. Furcron[7] essentially agrees with Keith. A. I. Jonas[8] and Jonas and G. W. Stose,[9] however, state that the Catoctin is the youngest pre-Cambrian formation in the Blue Ridge province in northern Virginia.

The writers have made a detailed study of the geology of the Blue Ridge in central Virginia between Buchanan, Botetourt County, and Afton, Nelson County. There are, in the area studied, two major outcrop areas of the Catoctin formation. This paper does not presume to offer conclusions based upon the geology of the entire Catoctin belt. Studies in central Virginia, however, have dis-

[1] Published with the permission of the state geologist of Virginia.

[2] *Geology of the Virginias* (New York: D. Appleton & Co., 1884), pp. 461–62.

[3] "The Volcanic Rocks of South Mountain in Pennsylvania and Maryland," *Amer. Jour. Sci.*, Vol. XLIV (3d ser., 1892), pp. 482–96.

[4] "The Structure of the Blue Ridge near Harper's Ferry," *Bull. Geol. Soc. Amer.*, Vol. II (1891), Pl. IV.

[5] "Geology of the Catoctin Belt," *U.S. Geol. Surv. 14th Ann. Rept.* (1894), Part II, pp. 306–18.

[6] Pp. 302–9, 311–15, 317–18, Pls. XXII and XXIII of ftn. 5.

[7] "Igneous Rocks of the Shenandoah National Park Area," *Jour. Geol.*, Vol. XLII (1934), pp. 405–7; and "Geology and Mineral Resources of the Warrenton Quadrangle, Virginia," *Va. Geol. Surv. Bull. 54* (1939), pp. 31–32.

[8] "Hypersthene Granodiorite in Virginia," *Bull. Geol. Soc. Amer.*, Vol. XLVI (1935), pp. 56–57.

[9] "Age Relation of the Pre-Cambrian Rocks in the Catoctin Mountain–Blue Ridge and Mount Rogers Anticlinoria in Virginia," *Amer. Jour. Sci.*, Vol. CCXXXVII (1939), pp. 579, 582–90.

closed evidence which seems to indicate that the Catoctin formation may be Cambrian in age.

If the stratigraphic relations of the Catoctin can be firmly established, the formation will provide a much-needed key to the interpretation of the geology of the Blue Ridge and Piedmont provinces in Virginia.

DISTRIBUTION

The Catoctin formation is exposed in the Blue Ridge province for about 200 miles from a point south of Carlyle, Pennsylvania, to a point in central Virginia several miles north of James River. Just north of Harpers Ferry the Catoctin belt bifurcates into two roughly parallel belts that trend southwestward into Virginia. The northwestern belt coincides with the summit area of the Blue Ridge. The southeastern belt is topographically expressed by a series of monadnocks, including Catoctin and Southwest mountains, which forms the foothills of the Blue Ridge. The Catoctin formation described in this paper crops out in the southern part of the northwestern belt (Fig. 1).

STRATIGRAPHIC RELATIONS

In the northeastern part of the Buena Vista quadrangle in central Virginia (Fig. 1), the Catoctin is underlain by a formation composed of conglomerate, arkosic sandstone, tuff, and andesite. Southwestward these rocks are metamorphosed and represented by quartzite, phyllite, and slate. This formation, to which the senior author in 1942 assigned the name "Oronoco"[10] from Oronoco Post Office, Amherst County, Virginia, unconformably overlies the pre-Cambrian crystalline basement complex of the Blue Ridge. This complex in central Virginia is composed essentially of a rock classed as hypersthene granodiorite.[11]

The Catoctin formation overlies the Oronoco formation and, where the Oronoco is absent, pre-Cambrian hypersthene granodiorite. Although the Catoctin is not distinctly or consistently stratified in the Buena Vista quadrangle, it is probably structurally parallel to the underlying Oronoco formation. This relationship is indicated by the parallel relationship, in rugged terrain, of contacts between the Catoctin and Oronoco formations and members in the Oronoco formation.

In eastern Augusta County in central Virginia, the lower portion of the exposed section of the Catoctin formation contains two members composed of tuff, conglomerate, and arkosic sandstone that are, respectively, 790 and 740 feet thick. The lithology of these members is essentially the same as the lithology of the Oronoco formation. Southwest of Augusta County no members comparable to those in Augusta County have been found in the Catoctin formation.

The Catoctin belt in Augusta County is thrust-faulted on both sides, so that neither the top nor the base of the formation is certainly exposed. The two members in that area, however, occur near the base of the exposed section. They may be at least partly equivalent to the Oronoco formation exposed to the southwest of the Catoctin section in Augusta County, or they may represent an upper part of the formation cut out by faulting farther to the southwest. The Catoctin may, in places, grade downward into a sedimentary formation

[10] Robert O. Bloomer, "Geology and Mineral Resources of the Buena Vista Quadrangle, Virginia," *Va. Geol. Surv.* (MS in files).

[11] Geologic map of Virginia, *Va. Geol. Surv.*, 1928.

(Oronoco) through the alternation of altered effusives and sedimentary members or upward into an overlying sedimentary formation (Unicoi) in the same manner. Exposures revealing these relations, however, do not occur in central Virginia.

The Catoctin in northern Virginia is overlain by the Lower Cambrian Loudoun formation,[12] which is at least in part equivalent to the Unicoi formation of central and southern Virginia. In central Virginia, however, the Unicoi rests upon the pre-Cambrian basement complex and not on the Catoctin. The Oro-

[12] *Ibid.*

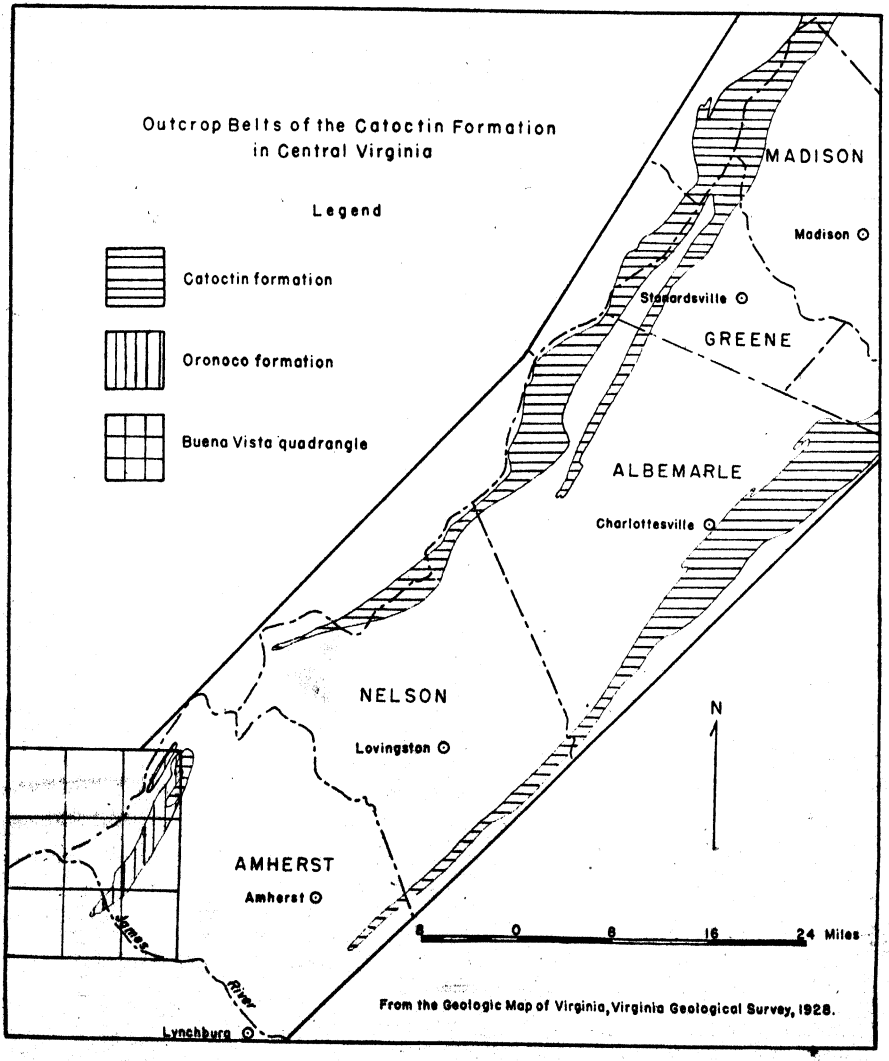
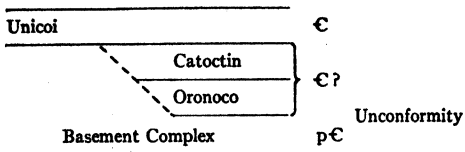


FIG. 1

noco formation, which underlies the Catoctin in central Virginia, rests on the pre-Cambrian crystalline complex except where it is overthrust upon the Unicoi. The Oronoco thins northwestward until the Catoctin rests directly on the pre-Cambrian crystalline complex. Presumably the unconformity beneath the Unicoi decreases in magnitude toward the southeast as the Catoctin and Oronoco come into the section. There is thus an overlapping relationship between the Oronoco, Catoctin, and Unicoi formations (Fig. 3).

TABLE 1

STRATIGRAPHIC RELATIONS OF THE CATOCTIN FORMATION IN CENTRAL VIRGINIA



STRUCTURE

At the southwestern end of the northwestern outcrop belt, in Amherst County, the Catoctin formation is exposed in an anticlinal block bordered on the southeast and on the northwest by overthrust faults. The fault along the northwestern edge of the Catoctin block passes beneath the fault along the southeastern edge just south of James River where the outcrop belt of the Oronoco formation is terminated. Granodiorite[13] forms the core of the anticlinal block, while the Catoctin and Oronoco formations are exposed on the limbs of the fold which plunges northward. A few miles north of James River the Catoctin outcrop belt ends, where it marks the nose of the faulted anticline.

Northeast of the Amherst County area, in Augusta County, the Catoctin is exposed in a block between the Blue Ridge fault and another overthrust fault. In this block, the Catoctin dips toward the southeast and thus represents a part of one limb of a large fold.

The Catoctin belt in central Virginia is bordered on the southeast by the Blue Ridge fault. This fault forms the northwestern margin of the Catoctin Mountain–Blue Ridge anticlinorium[14] which has consequently been thrust upon the Catoctin belt.

PETROGRAPHY

Megascopic character.—The Catoctin formation, as defined in this paper, is predominantly a fine-grained, holocrystalline, green, yellow-green, blue-green, gray, and red-brown altered effusive rock.[15] The structure is massive, except locally where it is slaty or mylonitic. The massive variety is complexly jointed and, in places, includes veins, veinlets, and podlike and irregular masses of milky or clear quartz, pink feldspar, and epidote. This vein material has undergone severe deformation in foliated varieties of the Catoctin formation. In these varieties, chlorite generally incases and obscures introduced aggregates of quartz, feldspar, and epidote.

The texture of the Catoctin propylite is generally uniformly fine grained, but porphyritic varieties with phenocrysts of altered euhedral feldspar occur locally. The texture observed in hand specimens is almost wholly secondary. It consists of an equigranular overgrowth of secondary minerals that obscures the primary tex-

[13] *Ibid.*

[14] Pp. 576–77 of ftn. 9.

[15] There are two members in the unit tentatively identified as the Catoctin formation in Augusta County. These are composed of sedimentary rocks and will be described in a detailed report being prepared for the Virginia Geological Survey.

ture. Incipient weathering clearly reveals the porphyritic character of some of the rock in which, on unweathered surfaces, the phenocrysts are obscured by secondary minerals.

Stratification is not clearly discernible in the Catoctin formation except where thin amygdaloidal and scoriaceous layers are present. It is presumed that the surfaces of these layers represent the surfaces of flows. The amygdaloidal and scoriaceous layers, unfortunately, are discontinuous and only sparsely developed. Hence they do not aid much in determining the attitude of the formation.

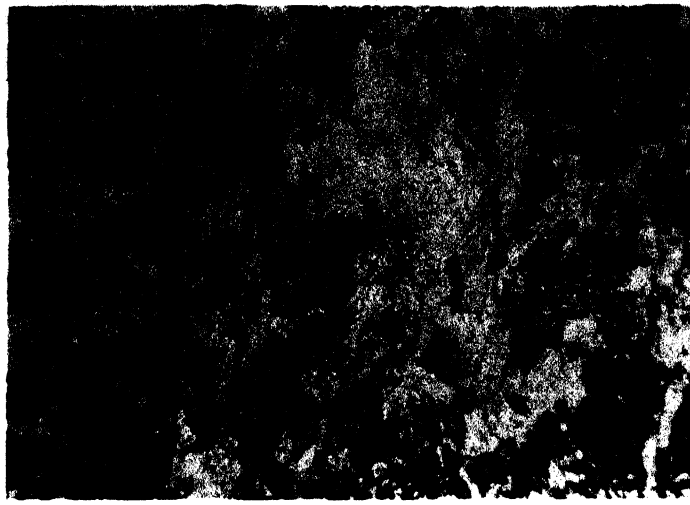
The amygdules in the Catoctin propylite range from about 3 to 15 mm. long. They are generally elliptical and contain pink feldspar, milky quartz, and epidote in all proportions. Other minerals in the amygdules include hematite, chlorite, chalcopyrite, malachite, and zeolite. In the slaty variety the amygdules are represented on S-planes by elliptical or ribbon-like smears dominantly composed of sericite and comminuted quartz.

With the exception of the minerals in the amygdules and rare phenocrysts, the only important rock-forming minerals recognized in hand specimens are epidote and chlorite. Minute grains of a black mineral, probably ilmenite, are discernible in some samples. Epidote appears to be very abundant. It forms an overgrowth that masks most of the primary grains and completely replaces others. The rock consequently appears monomineralic. In some specimens, chlorite can be detected in small micaceous grains subordinate to epidote.

Microscopic character.—Thin sections of the Catoctin effusive reveal a primary texture composed of laths of feldspar that average about 0.3 mm. in length. Phenocrysts of subhedral feldspar, rare in most specimens, range from 1 to 20 mm. in length. Small feldspar laths produce an ophitic fabric with the interstices occupied by fine-grained aggregates of secondary minerals. Secondary minerals also form discrete grains up to several millimeters in diameter, veinlets, saussuritic overgrowths, and embayments in primary grains.

Andesine, quartz, and orthoclase comprise the primary or magmatic minerals discernible in most thin sections. Andesine occurs in small laths. It forms about 50 per cent of the primary mineral composition of the rock. All the feldspar has been altered. Orthoclase forms tiny anhedral grains in the interstices and scanty subhedral phenocrysts. Many orthoclase grains have doubtlessly been completely replaced. Quartz forms minute grains associated with the secondary aggregate and rare larger grains in the interstices. Some of the quartz is secondary or introduced, but the largest grains are probably primary. Some aggregates of chlorite and epidote contain shredlike remnants of biotite. Scanty, minute grains of hornblende and pyroxene (augite?) have also been detected. It is believed, however, that the varietal femic mineral was biotite associated with accessory pyroxene. The hornblende is probably secondary. The recognition of biotite as the varietal mineral is strengthened by the occurrence of biotite in a comparatively thin, much less altered, effusive in the Oronoco formation 330 feet below the base of the Catoctin formation on the west side of Long Mountain in Amherst County. This effusive contains the same essential minerals as the Catoctin propylite. Furthermore, biotite comprises the femic constituent of an effusive in the base of the Unicoi formation (Cambrian) which is otherwise petrographically similar to the

PLATE I



A

B

A, Catoctin propylite showing an altered phenocryst of orthoclase (*O*), ghost-like laths of andesine, and a heavy overgrowth of secondary minerals. × *ca.* 100. *B*, Catoctin propylite showing vague outlines of andesine laths (*A*), quartz (*Q*) partially replaced by tremolite, and overgrowths of epidote, leucoxene, tremolite, and other secondary minerals. × *ca.* 100.

PLATE II

A

B

A, Propylitized andesite from the Oronoco formation. × *ca.* 100. *B*, Propylitized andesite from near the base of the Unicoi formation. × *ca.* 100.

Catoctin propylite. The effusives in the Oronoco and Unicoi formations are thought to be genetically related to the Catoctin propylite.[16] Some of the most abundant secondary minerals, such as sericite, epidote, chlorite, and iron oxide in the propylite were partly formed from biotite, of which tiny remnants remain. Biotite is also a prominent constituent of some of the dike rock in the granitic basement complex. These dikes are probably the conduits through which the Oronoco, Catoctin, and Unicoi lavas were supplied.

Some of the amygdules in the Catoctin propylite are filled with glass that contains spherulites and is devitrified on peripheries and along cracks. It is also partially replaced by secondary minerals, especially epidote and sericite. No glass has been identified with certainty in the groundmass of the rock. It may be present in minute grains, or it may have been destroyed by devitrification and replacement. The alteration products of the glass observed in amygdules are the same as those of andesine, orthoclase, and biotite and are, consequently, not distinctive.

With the possible exception of andesine, the most abundant minerals in the Catoctin propylite are secondary. These minerals occur as peripheral replacements and overgrowths on the primary minerals which generally appear as shadowy or ghostlike relicts in the groundmass of the rock. The primary features, except for amygdules and rare phenocrysts, are detectable only in thin sections. Even in some thin sections the primary characteristics are obliterated by secondary structures, textures, and minerals. The secondary minerals, in the approximate order of abundance, are: epidote, chlorite, sericite, tremolite, magnetite, ilmenite, hematite, calcite, quartz, leucoxene, and pyrite. The order of abundance and the proportion of the secondary minerals, however, are not invariable. The yellow-green propylite contains a preponderance of epidote. Chlorite is about as abundant as epidote in the blue-green variety. Tremolite is most abundant in the gray variety. A distinctive red-brown variety, so far observed only at the base of the formation in Amherst County, contains a preponderance of iron oxide, including, especially, specularite (Fig. 2).

CLASSIFICATION

Keith[17] recognized andesite, rhyolite, and basalt in the volcanic rocks that comprise the Catoctin formation in northern Virginia. Furcron[18] describes metabasalt, agglomerate, tuff, and sandstone among the rocks in the "Catoctin series" in the Warrenton area, Virginia. In central Virginia the Catoctin is an altered effusive thought to have been andesite or granobasalt. It is, however, difficult to recognize some of the primary characteristics of the rock because of the severe secondary alteration. The texture is clearly fine grained; andesine is the essential mineral; and biotite, orthoclase, and quartz seem to comprise the varietal minerals. Hence, the rock was probably andesite before it was altered.

The Catoctin formation is generally classified as "greenstone," a metabasalt.[19] This classification is somewhat confusing because it is also used to refer to soapstone, metagabbro, and other

[16] T. L. Watson and J. H. Cline, "Extrusive Basalt of Cambrian Age in the Blue Ridge of Virginia," *Amer. Jour. Sci.*, Vol. XXXIX (4th ser., 1915), p. 669.

[17] Pp. 302–9 of ftn. 5.

[18] Pp. 19–28 in *Bull. 54* of ftn. 7.

[19] See ftn. 11.

rocks. The Catoctin is essentially propylite. Perhaps this term should be used in preference to the more general term "greenstone."

Keith[20] also classified the Catoctin as a schist. In places the Catoctin is thinly foliated, but the writers have not seen specimens that could be classified as true schist. The foliated phase is too fine grained to permit the megascopic identification of essential micaceous minerals. Consequently, the foliated Catoctin propylite is perhaps more correctly classified as slate and, locally, as phyllite.

FIG. 2.—Ferruginous foliated phase of the Catoctin propylite showing laths of sericitized andesine. The opaque mineral is chiefly specularite. × *ca.* 100.

ORIGIN

The Catoctin formation in central Virginia is believed to be an altered plateau andesite that emanated from many fissures now represented by propylite dikes in the pre-Cambrian granitic complex of the Blue Ridge province. The rock in these dikes is like the rock in the Catoctin formation. None of these dikes has been found intruding the propylite flows. Lead-uranium ratios from the Catoctin formation and the more or less associated propylite, metagabbro, and metapyroxenite dikes, as well as from the effusives in the Oronoco and Unicoi formations, might disclose approximately equivalent ages and thus indicate a genetic relationship. If such a relationship were established, the widespread dikes and other bodies of propylite in the Piedmont and Blue Ridge provinces would assist greatly in the establishment of stratigraphic and age relations.

There is little doubt of the hydrothermal character of the solutions that altered the Catoctin andesite to propylite. The secondary minerals epidote, sericite, chlorite, calcite, and others and the absence of critical high-temperature minerals suggest that the altering solutions were epithermal. The source of these solutions, however, is problemati-

[20] P. 308 of ftn. 5.

cal. In the Blue Ridge and western part of the Piedmont province, where the Catoctin propylite is extensively exposed, there are no known large, intrusive masses of post-Catoctin age that could have supplied the solutions that altered the formation. Furthermore, with a few local exceptions, the rocks upon which the Catoctin formation rests, including arkosic sandstones, siltstones, tuffs, and granitic rocks, are not similarly altered. No veins or dikes that could have supplied the solutions that so extensively propylitized the Catoctin formation have been found in these underlying rocks. Hence, it is probable that the altering solutions were deuteric[21] solutions in the Catoctin andesite. These solutions probably were retained within the flows by the development of chill-crusts on the lower and upper surfaces. Locally, where cracks were formed in the basal crust, solutions may have escaped and produced hydrothermally altered zones in the underlying rocks. In places, lithified zones in the flows may have been broken or fractured by the movement of the molten effusive. These fracture zones within the flows may have become channels which, when filled with deuteric solutions, formed the veins and masses of quartz, orthoclase, and epidote observed in the Catoctin formation. Some of these veinlike masses containing approximately equal amounts of coarse-grained pink feldspar and quartz resemble pegmatite and granite. The field relations, mineral assemblage, and paragenesis, however, indicate the hydrothermal origin of these masses. Epidote is associated in all proportions with the quartz and feldspar and is contemporaneous with these minerals. There are, furthermore, no critical magmatic minerals. The same mineral assemblage is found in the amygdules.

[21] P. 315 of ftn. 5; and p. 24 in *Bull. 54* of ftn. 7.

THICKNESS

The thickness of the Catoctin formation has not been accurately measured. Such a measurement is difficult, if not impossible, because the exposed sections are faulted at the top or bottom, or at both the top and bottom, and no flow boundaries can be identified with certainty within the propylite. In Amherst County where the base of the section is exposed but the top faulted, the Catoctin is represented by about 2,200 feet of propylite. Northward, in Augusta County, the Catoctin is exposed between faults so that the section is incomplete. There are, however, in the Augusta County section, about 5,000 feet of propylite and sedimentary rocks. The sedimentary rocks occur in the lower part of the section where they form two mappable units (members?), respectively 740 feet and 790 feet thick. In northern Virginia, Keith[22] assigned a thickness of only 1,000 feet to the Catoctin formation. Furcron[23] described an agglomerate

TABLE 2

GENERALIZED SECTION OF THE CATOCTIN FORMATION FROM EXPOSURES IN AMHERST COUNTY, VIRGINIA

	Thickness (In Feet)
Unicoi formation	
Catoctin formation (2,200 feet, estimated)	
4. Propylite, uniformly fine grained, green, yellow-green, gray; recurrently amygdaloidal, massive; locally foliated	1,550
3. Propylite, porphyritic, fine grained, green, massive	100
2. Propylite, fine grained, green, massive	500
1. Slate, red, light-green; flattened, sericitized amygdules	50
Oronoco formation and granodiorite	

[22] Pl. XXIII of ftn. 5.

[23] P. 22 in *Bull. 54* of ftn. 7.

member at least 1,500–2,000 feet thick at the base of the Catoctin in Fauquier County in northern Virginia. Furcron, however, gives no estimate of the total thickness of the formation.

AGE

Keith[24] expressed the opinion that the Catoctin formation is the oldest rock in the Blue Ridge of northern Virginia. He stated that the Catoctin is intruded and metamorphosed by pre-Cambrian granitic rock (granite and granodiorite) but offered no concrete proof of the post-Catoctin intrusion of pre-Cambrian plutons.

Furcron[25] agrees with Keith with regard to the pregranitic complex age of the Catoctin. He reached this conclusion because of an intrusion of "pegmatite" in the Catoctin and because of exposures of "numerous alternations of granite and greenstone [Catoctin] layers." The "pegmatite" mentioned by Furcron is assumed by him to be genetically related to the pre-Cambrian granitic complex in the Blue Ridge. However, as previously stated, these pegmatite-like bodies are probably epithermal in origin and genetically related to the andesite effusives from which the Catoctin formation was formed. Some of these bodies containing quartz and orthoclase in variable proportions closely resemble some pegmatites or granites. Epidote is present in nearly all these bodies and seems to be contemporaneous with the quartz and feldspar. Furthermore, such bodies have not been found in rocks older or younger than the Catoctin formation in central Virginia. These observations and the absence of mica and critical high-temperature minerals suggest that the so-called "injections" are epithermal veins and masses formed from residual solutions in the Catoctin effusives. With regard to the "alternations of granite and greenstone layers," Furcron assumed that the "granite" is the younger rock but states that it has not been found cross-cutting the propylite ("greenstone"). He recognized, however, a body of marble, slate, schist, and gneiss beneath the Catoctin to which he assigned the name "Fauquier formation."[26] Hence, Furcron does not recognize the Catoctin formation as the oldest rock in the pre-Cambrian complex of the Blue Ridge region of northern Virginia as did Keith. The relation of the Fauquier formation and the pre-Cambrian granitic complex is not described.

Jonas[27] and Jonas and Stose[28] are of the opinion that the Catoctin is the youngest pre-Cambrian formation in the Blue Ridge. This conclusion is based upon the absence of pre-Cambrian intrusions in the Catoctin formation, the presence of clastic sediments in the Catoctin derived from the granitic complex, and the extrusion of the Catoctin upon the eroded surface of the granitic basement complex.

The stratigraphic relations and lithology of the Oronoco, Catoctin, and Unicoi formations in central Virginia are indicative of the Lower Cambrian age of these formations.

The Catoctin overlies the Oronoco formation and (where the Oronoco is missing) the pre-Cambrian granitic complex (granodiorite). The Oronoco formation contains conglomerate, tuff, arkose sandstone, and an andesite effusive. Pebbles, cobbles, and boulders of granodiorite

[24] Pp. 302, 309, 311–15, 317–18, Pls. XXII and XXIII, of ftn. 5.

[25] Pp. 405–10 in *Jour. Geol.*, and pp. 31–32 in *Bull. 54* of ftn. 7.

[26] Pp. 37–41 in *Bull. 54* of ftn. 7.

[27] Pp. 56–57 of ftn. 8.

[28] Pp. 579–90, 592 of ftn. 9.

from the pre-Cambrian granitic complex form the conglomerates in the Oronoco formation which concordantly underlies the Catoctin formation. The Oronoco is thus younger than the pre-Cambrian complex and older than the Catoctin formation.

In central Virginia, no rocks resting upon the Catoctin formation are exposed. Farther north, however, the Catoctin is unconformably overlain by the Loudoun formation.[29] In central Virginia the Unicoi formation, at least in part equivalent to the Loudoun formation, is not underlain by the Catoctin but rests unconformably on the granitic complex. The basal portion of the Unicoi contains tuff, an andesite effusive (locally propylite), and a conglomerate composed of cobbles and pebbles from the granitic complex. Propylite fragments are extremely rare or absent in this conglomerate. If the absence of the Catoctin formation beneath the Unicoi formation is due to erosion, the clastic rocks in the Unicoi should, it seems, contain fragments of propylite. It is consequently assumed that, in central Virginia, the Oronoco and Catoctin formations were developed mainly in a basin east of the area of the present Unicoi outcrop belt along the west flanks of the Blue Ridge and never spread westward over the rocks that now underlie the Unicoi. Such an Oronoco-Catoctin basin to the east would explain the absence of the Catoctin beneath the Unicoi formation but would not account for the sparsity of propylite fragments in the Unicoi conglomerates. The erosion of the Catoctin and westward transportation of the sediment would have supplied the Unicoi with propylite fragments. Hence, at least during the first stages of deposition, the Unicoi sediments may have come from a crystalline area to the west of the present outcrop belt of the Unicoi in central Virginia. This area may now be buried beneath Paleozoic sediments in the Appalachian trough west of the Blue Ridge. On the other hand, overthrust faults have moved the present outcrop belt of the Catoctin westward so that it is exposed much nearer the present outcrop belt of the Unicoi formation than it was during the time of Unicoi deposition There may have been an intervening area, not covered by Catoctin propylite, sufficiently wide to permit much dilution and attrition of the Catoctin-derived sediment. This may account for the sparsity of propylite fragments in the Unicoi conglomerates exposed along the west flanks of the Blue Ridge in central Virginia.

The clastic sediments of both the Unicoi and Oronoco formations were derived from the pre-Cambrian granitic complex and are very smilar in appearance. Furthermore, the tuffs and andesites in the two formations are very similar in color, texture, mineral composition, and alteration. In fact, without the aid of field relations it is difficult, if not impossible, to distinguish the rocks in the Oronoco formation from those in the lower portion of the Unicoi formation.

The Unicoi is conformably overlain by the Hampton formation, which in turn is conformably overlain by the fossil-bearing Lower Cambrian Erwin formation. This relation points to the Cambrian age of the Unicoi formation.

The Catoctin seems to have been formed from an andesite of the same composition and texture as the andesites in the Oronoco and Unicoi formations. This and the presence of lithologically similar red and gray tuffs in the Oronoco and Unicoi formations are indicative of a period of volcanism that began during

[29] See ftn. 11 and pp. 42–43 in *Bull. 54* of ftn. 7.

Diagrammatic Representation of the Stratigraphic Relations of the Unicoi, Catoctin and Oronoco Formations

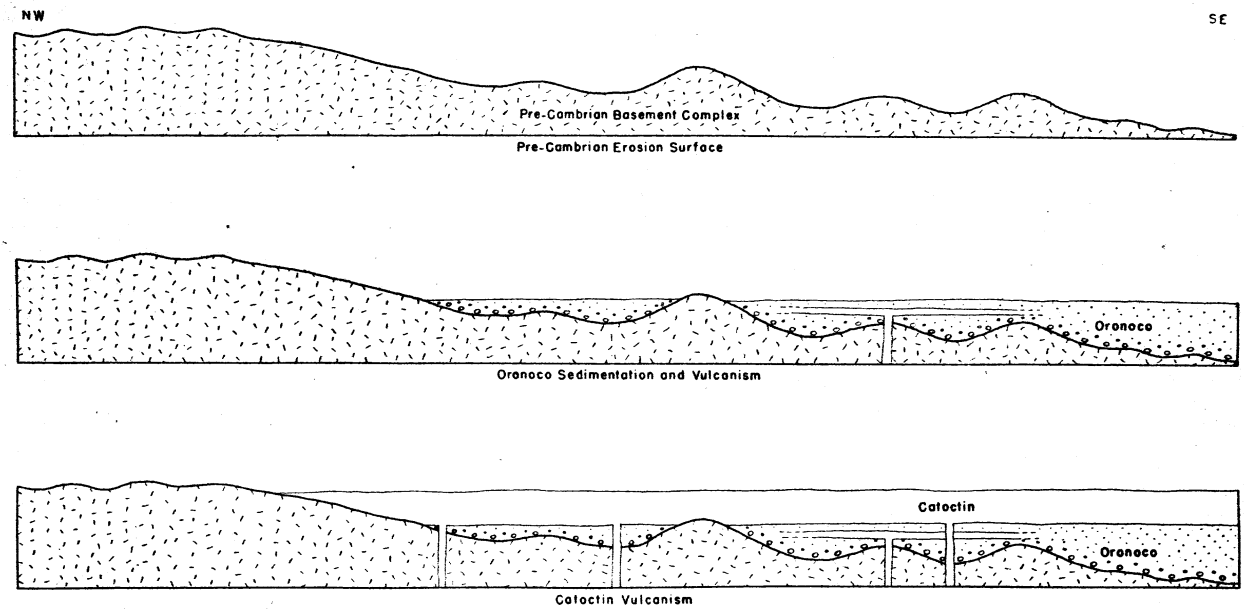
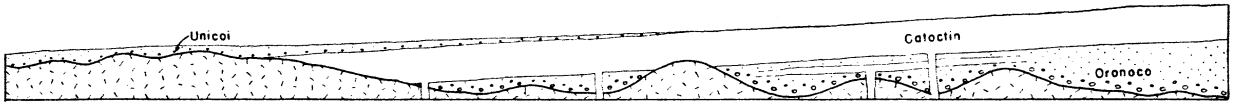
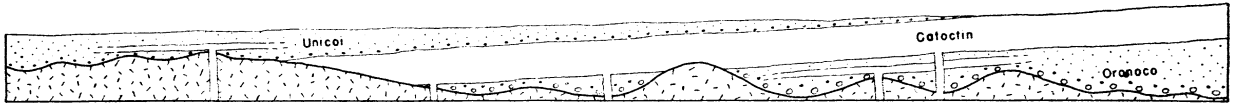


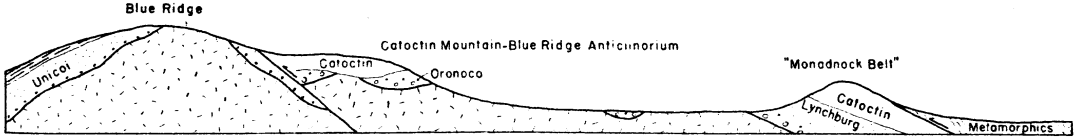
FIG. 3



Westward Tilting and Unicoi Sedimentation Initiated



Unicoi Vulcanism and Sedimentation



Idealized Structure Section of the Blue Ridge North of James River
Showing Present Topography

Oronoco time, continued throughout Catoctin time, and waned during the early part of Unicoi time.[30] Perhaps this volcanism accompanied the initial crustal deformation that resulted in the formation of the Appalachian geosyncline and is consequently one of the first events of Paleozoic time.[31]

The similarity of the lithology and the possible genetic relationship of the volcanics in the Oronoco, Catoctin, and Unicoi formations suggests that these formations were formed under similar conditions. Each of these formations rests upon the eroded surface of the pre-Cambrian granitic complex in the Blue Ridge. In places the Catoctin formation is separated from this significant unconformity by the Oronoco formation, but the Unicoi in central Virginia rests directly upon it. The absence of the Catoctin and Oronoco formations beneath the Unicoi formation in central Virginia probably is due to the relief of the erosion surface on the pre-Cambrian granitic complex (Fig. 3).

The relations and the lithologic similarity of the Oronoco, Catoctin, and Unicoi formations lead to the conclusion that these formations are a continuous stratigraphic series. If the Unicoi formation is Cambrian, then the Catoctin and Oronoco formations are also Cambrian.

The Oronoco formation has not been identified beyond the confines of the Buena Vista quadrangle in central Virginia. Its relation to other sedimentary and metasedimentary formations beneath the Catoctin is problematical. Jonas and Stose,[32] however, have stated that the Lynchburg gneiss (metasediment) is part of the younger pre-Cambrian series that contains the Catoctin formation in the upper part. This suggests that the Oronoco formation, which is overlain by the Catoctin, may be at least partly equivalent to the Lynchburg formation. The Lynchburg formation may then be Cambrian in age. This formation and the Catoctin formation are extensively exposed along the western margin of the Piedmont province in Virginia. The establishment of the age of these formations relative to known Paleozoic formations would greatly aid in interpreting the geology of the Piedmont and Blue Ridge provinces.

ACKNOWLEDGMENTS.—These geologic studies of the Blue Ridge in central Virginia were made for the Virginia Geological Survey. The writers appreciate the suggestions and aid of members of the survey staff. They are especially grateful to Arthur Bevan, W. R. Brown, R. S. Edmundson, and Mr. and Mrs. G. W. Stose for critically reading the manuscript.

[30] See ftn. 16.

[31] G. F. Laughlin, "Comments on the Origin and Major Structural Control of Igneous Rocks and Related Mineral Deposits," *Econ. Geol.*, Vol. XXXVI (1941), p. 674.

[32] Pp. 589–90 of ftn. 9.

A FAULT-PLANE CAVERN

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ABSTRACT

The Goodwins Ferry Cave is a linear horizontal cavern having two accessible levels. The linear nature of the cave is due to development on the plane of the Saltville fault, while an earlier control of the drainage by the New River caused the cave to develop horizontally. Enlargement has taken place by solution and caving, and the cave's water is thought to come from overlying sandstones.

LOCATION

The Goodwins Ferry Cave[1] is located in the folded Appalachian Mountains near Goodwins Ferry in Giles County, Virginia, about 25 miles northwest of Blacksburg. The entrance to the cave is at an elevation of 2,015 feet on the end of Spruce Run Mountain, sometimes called Cave Hill, and about 235 feet above the road through the New River gorge.

GEOLOGY OF THE CAVE AREA

I. STRATIGRAPHY

The rocks exposed at Spruce Run Mountain, in the vicinity of Goodwins Ferry, include beds of Ordovician and Silurian age against which Cambrian formations have been thrust along the well-known Saltville fault. The columnar section of these strata is shown in Table 1.

LOWER CAMBRIAN

Honaker dolomite.—The Honaker formation is a thick-bedded, fine-grained, grayish, nonfossiliferous dolomite. No limestone was found at any of the outcrops. The dolomite in some places has been moderately fissured, but the fissures are tightly sealed by veins of dolomite crystals. Near the fault the beds are crenulated with minor folds.

Nolichucky shale.—The Nolichucky formation, of fossiliferous shale and dolomitic limestone, is well exposed about $\frac{1}{4}$ mile south of Goodwins Ferry along a cut of the Virginian Railway. Here a gradational contact with the Honaker dolomite is exposed. The Nolichucky shale dips south at about 55°.

TABLE 1

System	Formations	Thickness (Feet)
Silurian	Clinton ss	80
	Cacapon ss	120
	Clinch ss	55
Ordovician	Juniata sh	95
	Martinsburg sh-ls	800
	Eggleston ls	400
	Moccasin ls	60
	"Lowville" ls	150
	Benbolt ls	100
	(SALTVILLE FAULT)	
Lower Cambrian	Nolichucky dol-sh	55
	Honaker dol	800

ORDOVICIAN

Benbolt limestone.—The Benbolt formation is composed of a dark-blue, fossiliferous limestone alternately thin- and moderately thick-bedded. The most outstanding characteristic is that the beds are made up of small nodules of limestone 1–2 inches in diameter. These nod-

[1] The cavern is generally called the Goodwins Ferry Cave because of the near-by settlement, but it is also known as the New River Cave and, infrequently, as Spruce Run Cave.

ules are believed to result from the jointing and weathering of thin layers of limestone having thin partings of calcareous shale.[2] The Benbolt limestone is not very well exposed on the Goodwins Ferry side of the river, but on the opposite slope, along the cut of the Norfolk and Western Railroad, there is an excellent outcrop showing beds intensely folded near the fault contact with the Honaker dolomite. On the Cave Hill, near the contact with the Honaker dolomite, the limestone has been deformed only by gentle, wavy folding. The contact with the Martinsburg shale represents faulting, but the fault is obscured by soil mantle, and its true nature is not readily determinable. About ½ mile north of the cave the formation has excellent outcrops on both sides of the New River gorge and shows folding with the overlying "Lowville."

Identification of the Benbolt limestone was made by Dr. Byron N. Cooper on fossil evidence collected by the writer. Heretofore, the presence of the Benbolt formation on the Saltville fault contact has not been recognized, and it is the author's belief that the formation here has always been erroneously included with the overlying dolomite. As a result, the plane of the Saltville fault has been mistakenly placed along a minor fault associated with the major thrust.

"Lowville" limestone.—The "Lowville" formation, also best exposed along the railway cuts, is a dark bluish-gray, fine-grained, compact, nonfossiliferous limestone. The formation is generally thin- to medium-bedded, with the thicknesses averaging between 1 foot and 2½ feet, although a few beds run around 6–8 inches. There are several chert horizons near the top of the formation.

[2] Charles Butts, "Geology of the Appalachian Valley in Virginia," *Va. Geol. Surv. Bull. 52* (1940), Part I, p. 171.

Moccasin limestone.—The Moccasin formation is dark-pink, though in places gray-buff, argillaceous limestone, in which deformation has developed pronounced fracture cleavage that gives its outcrops a slatelike appearance. The formation is compact and medium-bedded, the beds averaging from 1 foot to 1½ feet in thickness. The formation is well defined, continuous, and somewhat resistant to weathering, forming a slight ridge in places.

Eggleston limestone.—Moderately folded with the Moccasin limestone is the overlying Eggleston formation, which is a thin- to medium-bedded, fine-grained, argillaceous, buff limestone characterized by a basal chert conglomerate and cuneiform jointing perpendicular to the bedding. The formation contains several beds of bentonite, and the peculiar jointing may be related to this material. The basal conglomerate suggests a disconformity.

Martinsburg shale.—The Martinsburg formation is predominantly a yellowish to brownish shale with numerous limestone beds 1–4 inches in thickness. Very thick beds of clay shale are numerous and quite fossiliferous. Where the shale is thin-bedded, it weathers to an easily recognizable flaky talus. The only good outcrop noted is along the road cut, ½ mile north of Goodwins Ferry, and here only a small part of the formation is exposed. Elsewhere the shale is covered with a soil mantle that obscures it almost completely.

Juniata shale.—The Juniata formation is composed mainly of bright-red shale and beds of brown to red sandstone. The shale is compact, unfossiliferous, and covered with a soil mantle so that there are very few outcrops.

SILURIAN

Clinch sandstone.—The Clinch is a sandstone composed almost entirely of quartz sand and is generally light gray to white. It is a medium-textured, compact, vitreous rock that is practically a quartzite. Its resistant nature determines numerous outcrops mostly as ledges on steep slopes. The sandstone is thick-bedded, with many ripple-marked bedding planes. At the base of the formation, for a thickness of some 30 feet, are conglomeratic layers containing quartz pebbles up to $\frac{3}{4}$ inch in length.

Clinton sandstone.—The Clinton formation is also a compact, thick-bedded, and well-jointed sandstone. The lower half of the formation has highly ferruginous sandstone beds of distinctive character, termed the Cacapon division. The Cacapon sandstone is moderately thin- to thick-bedded and has a striking purplish to brownish-red coloration owing to its iron content. The Clinton formation above the Cacapon division is predominantly a white to gray, medium-textured sandstone having plentiful outcrops, especially along steep slopes.

II. STRUCTURE

The major structural feature of the Goodwins Ferry area is the Saltville or Bland fault extending 200 miles from the north end of Sinking Creek Valley (Craig County, Va.) southwest through Bland and Saltville, and well into Tennessee. The maximum stratigraphic displacement occurs in Washington County, where the Lower Honaker of Cambrian age is brought up against the Mississippian Pennington formation—a stratigraphic distance of 14,000 feet.[3]

On Spruce Run Mountain the Lower Cambrian dolomite is in contact with the Silurian sandstones with a stratigraphic displacement of 4,800 feet. The Honaker dolomite has been thrust from the southeast up against the section which includes the Ordovician and Silurian beds already described. These beds have been bent into a syncline, and the southern limb of the fold has been pinched off by a minor fault which has brought a sliver of the Benbolt limestone into contact with the other Ordovician beds as high as the base of the Clinch sandstone (Fig. 1). The syncline appears to be overturned toward the north and is most intensely folded near the fault.

The Saltville fault, on the slope above the cave entrance, has a dip of 30°, which, below the entrance, rapidly increases to around 50° and is 54° at the foot of the hill. The strike of the fault is, in general, N. 65° E. The minor fault probably has the same strike, but its dip is more uniform, being around 52°.

Along the plane of the Saltville fault the beds of the Honaker dolomite dip southward at the same angle as the fault and then curve upward again to form a syncline whose axis is in the Goodwins Ferry hollow and parallel to the strike of the major fault. South of the syncline the dolomite is in the form of an anticline of about the same magnitude as the syncline and is overlain on the south limb by the Nolichucky shale.

THE GOODWINS FERRY CAVE

I. CAVE DESCRIPTION

The Goodwins Ferry Cave has two predominant levels (Fig. 2). The older, upper level is, in many places, 75 feet in maximum width, while the lower level, where the stream flows, is comparatively small as befits its recent development.

The entrance to the cave is on the fault plane, and both levels of the cave extend

[3] *Ibid.*, p. 457.

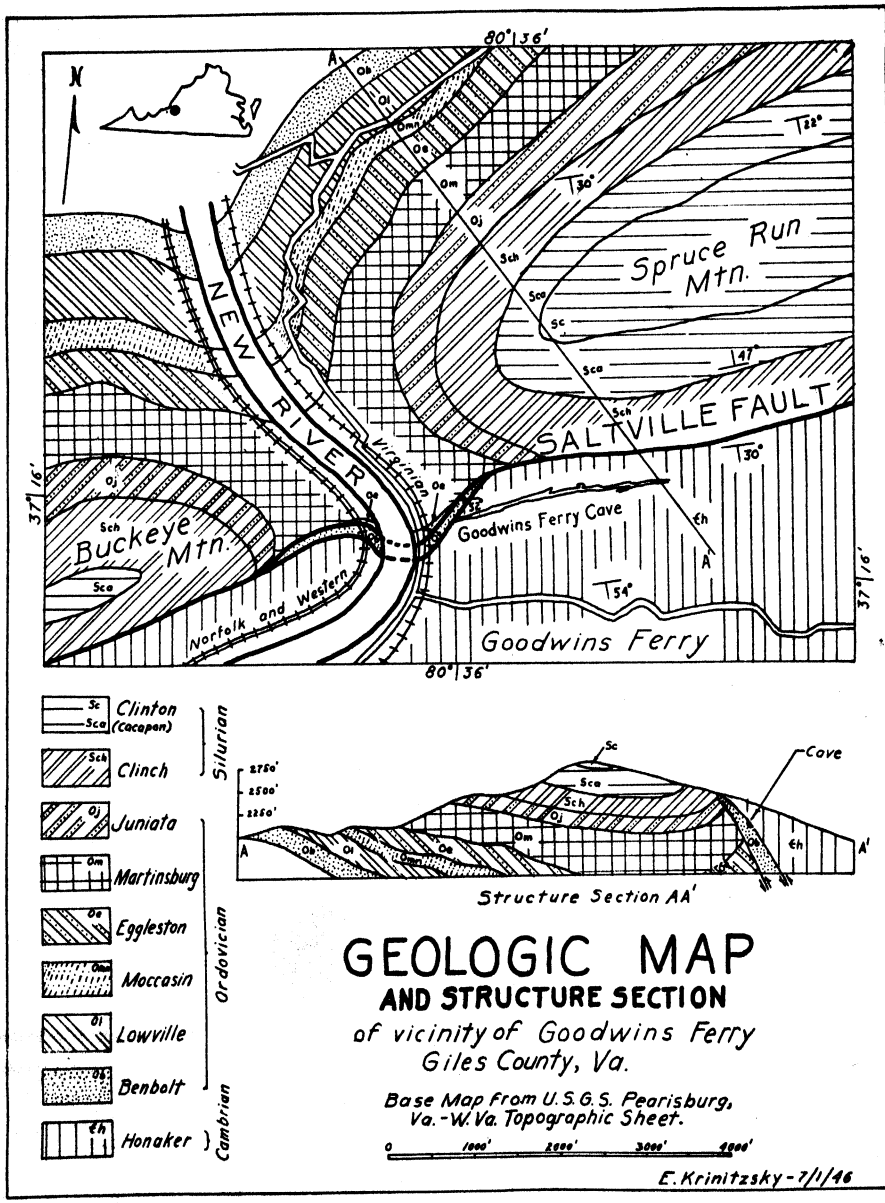
80°|36'
N
Spruce Run Mtn.
NEW RIVER
SALTVILLE FAULT
Virginian
Goodwins Ferry Cave
Buckeye Mtn.
Norfolk and Western
Goodwins Ferry
37°|16'
Sc Sca Clinton (Cacapon)
Sch Clinch
Oj Juniata
Om Martinsburg
Oe Eggleston
Om Moccasin
Ol Lowville
Ob Benbolt
Silurian
Ordovician
Cambrian
Cave
Structure Section AA'
GEOLOGIC MAP
AND STRUCTURE SECTION
of vicinity of Goodwins Ferry
Giles County, Va.
Base Map From U.S.G.S. Pearisburg, Va.-W. Va. Topographic Sheet.
0 1000' 2000' 3000' 4000'
E. Krinitzsky - 7/1/46

FIG. 1

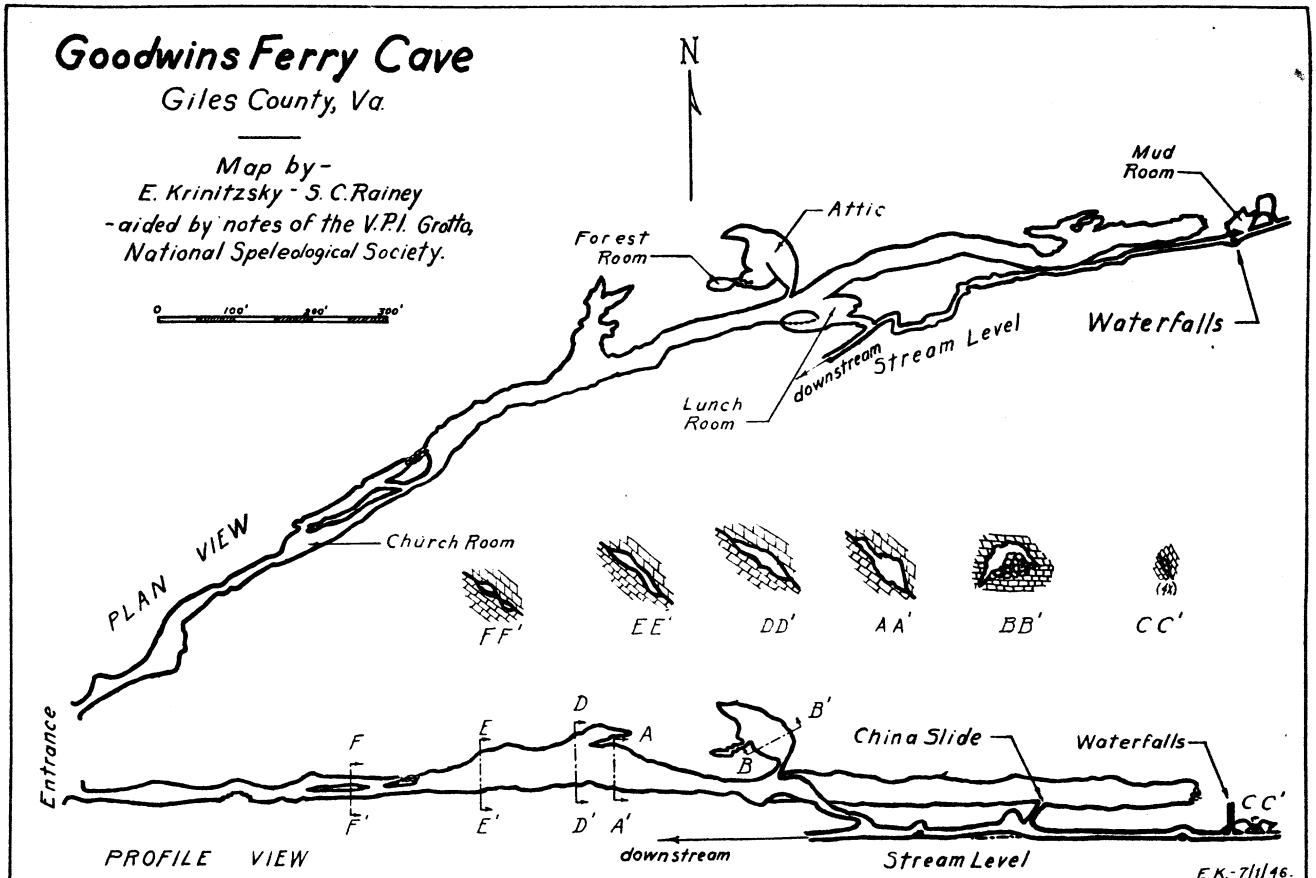
Goodwins Ferry Cave
Giles County, Va.
Map by -
E. Krinitzsky - S. C. Rainey
- aided by notes of the V.P.I. Grotto,
National Speleological Society.
N
0 100' 200' 300'
Mud Room
Attic
Forest Room
Waterfalls
Stream Level
downstream
Lunch Room
Church Room
PLAN VIEW
FF'
EE'
DD'
AA'
BB'
CC'
Entrance
F
F'
E
E'
D
D'
A
A'
B
B'
China Slide
Waterfalls
CC'
PROFILE VIEW
downstream
Stream Level
E.K.-7/1/46.

FIG. 2

along the strike of the fault for ½ mile. The fault dip, which averages 53° through most of the cave, dominates the shape of the passageways (Fig. 2). Sections *EE′* and *DD′* show characteristic cross sections. The stream has worked downward along the fault surface to form a tabular-shaped room. In section *AA′* the same type of stream action has been working down the dip, but large rock-falls have caused an enlargement upward. Debris from the rock-falls has been removed by the stream. Section *BB′* shows enlargement in what was probably a zone of weakness due to pronounced jointing. Single blocks, as large as 50 × 50 feet, have fallen from the roof and have built up a huge mound that apparently could not be removed by whatever water circulation there was at the time of the room's development. This room, known as the "Attic," is not the only one that owes its incongruous shape to lines of weakness other than the fault plane. Similar enlargements are the "Mud Room" and the "Waterfalls." In section *CC′* there is shown a striking drop of the stream level in a recent part of the cavern.

Enlargement of the cavern by stream action appears to have been done almost wholly by solution, and, although the cavern has developed along the fault plane where dolomite is in contact with limestone, solution by the stream has caused enlargement chiefly in the limestone. In the front rooms of the cave the roof and right-hand wall (facing inward) are of dolomite, whereas the floor and left-hand wall are of limestone. At the rear of the first room is an accumulation of fallen blocks, which average 6 × 5 × 3 feet, and all of which appear to be of dolomite. In the split of the passageways, near the "Church Room," there are immense numbers of platelike dolomite blocks. They are roughly circular, and many are from 5 to 15 feet in diameter but are only 10 inches to 3 feet in thickness. Farther inward the roof is still of dolomite, but the floor slope is much obscured by flow stone. Wherever bedrock outcrops, however, it is everywhere of limestone. In the Attic the fallen blocks are all of dolomite. Even on the stream level the same association of limestone floor with dolomite roof and dolomite rock-falls can be seen. From the foregoing it may be said that all the rock-falls are probably of dolomite and that enlargement of the cavern into the dolomite has proceeded solely in this manner.

The extremely linear development of the cave (Fig. 2) was caused by its situation on the fault plane, but the horizontal nature of the passageways calls for some other type of control in their development.

II. HYDROLOGY

Shortly after the cave stream ceased flowing in the upper level of the cave, the outlets of the center rooms became dammed, and ponded waters formed good-sized lakes which are now evidenced by flowstone rims that once marked their periphery.

The upper levels of the cave are at present generally dry except for small trickles in the middle rooms which aid dripstone deposition and which form a few tiny pools even now. It may be that at an earlier time there were heavier seepages, the water of which has been diverted or the fissures sealed with dripstone. Prolonged rainy weather moderately increases the number and volume of the trickles and the general dampness of the cave so that there is a greater possibility that the earlier flow of water has been diverted from the fissures.

Dolomite rock-falls are common on the present stream level, but they never have the effect of damming the stream. The water readily percolates through all the rock-fall accumulations that strew the stream's path, and so it can be reasoned that caving of the dolomite roof could not be solely responsible for the damming that took place in the upper level. Most likely the water that seeped into the room was heavily charged with dissolved materials and, by depositing dripstone, served to seal off rock-fall accumulations and so in itself created the barriers to its movement.

At the lowest level there is a permanent stream which flows westward to the New River. In places the stream cross section is roughly $3\frac{1}{2}$ feet wide and $1\frac{1}{2}$ feet deep, but in general the stream flows over a wider, rocky bed. At the Waterfalls (Fig. 2) there is a chimney with a vertical height of 80 feet and a circular cross section around 11 feet in diameter. A heavy fall of water furnishes roughly a third of the volume of the stream.

At one time the stream flowed through the upper level and out of the present cave entrance and there cascaded down the slope as the stream now does at the lower level. There yet remains directly below the entrance a pronounced hollow filled with rocky debris through which the stream originally flowed. At present there is a built-up ledge at the entrance, over which one descends to enter the cavern. No doubt this ledge is the result of the accumulation of rock and soil talus washed down from the hillside above since the stream abandoned this exit.

On the Cave Hill, 75 feet above the road, the cave stream emerges through rock talus as a spring. This emergence is at the level of the stream within the cave, but it accounts for only about half the volume of the cave stream. Below the road level and 70 feet to the south of the cave spring, another spring issues from talus, its volume perhaps somewhat increased by near-surface circulation. It contains the remainder of the cave water. Fluoroscein tests made on the cave waters showed distinct coloration in both springs. The downstream level of the cave can be followed almost to the outlet of the upper spring, but the points at which the water turns down to the lower spring cannot be determined with exactness, as the water is lost gradually.

Pebbles of quartz, limestone, and dolomite occur along the stream bed. Most abundant are those of dolomite, which are sharply angular and are nearly all solution-pitted, especially along lines of joint or fracture weakness. The quartz pebbles are all well rounded and, in all probability, are from the basal conglomerate in the overlying Clinch sandstone and were not rounded by the action of the cave stream. The presence of these quartz pebbles from the Clinch conglomerate indicates a water circulation from the overlying sandstones toward the cave.

The cave stream is remarkably regular in flow, being little affected by rainstorms, and the absence of any sinkhole pattern or major sinkholes on the surface further indicates that the water in the cave must come from the sandstones. But, from the vitreous and compact nature of the sandstones, it seems unlikely that they could act as a porous reservoir. Jointing, however, is fairly well developed in the Clinch sandstone, and seepage along such fissures may be in considerable volume. The synclinal structure of the Juniata shale (Fig. 1), by acting as a basin and keeping a perched water table in the sandstones, could furnish an over-

lying source of water for the cavern. Rainfall in Giles County averages 39 inches per annum and is fairly evenly distributed.[4] Monthly precipitation is around 3.2 inches, with a high of 4.28 inches in July and a low of 2.44 inches in October, the remaining months showing fairly uniform rainfall. It is highly probable that the uniform rainfall along with slow seepage through the sandstones and from the perched water table are responsible for the steady flow of the cave stream.

Fine muds and sands are absent in the stream channel except for small quantities associated with pebble accumulations in hollows. Where the stream beats against fallen blocks of dolomite, there is considerable indication of solution hollowing in the form of pits and grooves on the rocks. Thus it seems likely that abrasion plays but a small part in the stream's erosion and that solution is the principal factor.

III. DEPOSITS

The first two rooms from the entrance are bare except for an occasional nub of a stalagmite, but beyond, on the same level, the rooms have a complete veneer of beautiful flowstone with many elephant's-ear pendants, magnificent stalagmite masses, and several columns. The Lunch Room, the Mud Room, and the stream level in between are of comparatively recent origin and appropriately rugged and bare of dripstone. The stream level, downstream from the Lunch Room, has several spots containing choice deposits, but the best development in the entire case is in the Forest Room.

Although there are many small helictites in the main level, these interesting forms are best developed in places on the ceiling of the Attic. The helictites of the Goodwins Ferry Cave (Fig. 3) are characteristically elongated spines of aragonite which radiate outward in all directions. The spines are generally straight and very slender, although in places the needles of aragonite bunch themselves together to form a stout rod which may be slightly curved. Secondary developments of the aragonite needles radiate outward from growth centers on primary stalks. At many points in the cave are found the common aragonite knobs, but in the Attic there is an unusual form in which the knobs appear to have been pushed out on several stalks.

IV. ORIGIN

Since Davis presented his exhaustive treatment of the two-cycle theory of cave origin,[5] it, and subsequent theories by other authors, has focused attention on the extent to which a cave has been phreatic. Davis believed that most caverns originated below the water table and were completely filled with water during the time of their enlargement; later he postulated a change of conditions which lowered the water table and so drained the caves. Then, he thought, a second cycle began, in which cave enlargement either ceased or proceeded at a much slower rate and during which time dripstone deposition began.

If their very beginning is considered, all caves are phreatic in origin, for the first circulation of water in some incipient fissure that originally determined the cave must have completely filled the tiny course. When a cave is below the water table or in a saturated zone, huge rooms[6]

[4] *Climate and Man: Yearbook of Agriculture* (Washington: U.S. Department of Agriculture, 1941), p. 1159.

[5] W. M. Davis, "Origin of Caves," *Bull. Geol. Soc. Amer.*, Vol. XLI (1930), pp. 475–628.

[6] B. C. Moneymaker, "Subriver Solution Cavities in the Tennessee Valley," *Jour. Geol.*, Vol. XLVIII (1941), pp. 74–86.

can develop from small circulations, and such rooms will be completely filled with water. However, if the cave is not in a water-saturated horizon, the cave void cannot be at all times completely filled by the cave stream unless the volume of the cave is equal to or less than the volume of the stream at its normal rate of flow. When such a stream enlarges its path to a point where the cavern is larger than the volume of the stream, the cave ceases to be phreatic. But enlargement should not slow down at this time. As long as the stream does not leave the enlarged cave course, and as long as the rate of circulation does not diminish, the stream will continue to excavate as it did previously, and, at the same time, dripstone may be deposited. The author believes that such conditions prevailed in the development of at least the lower level in the Goodwins Ferry Cave.

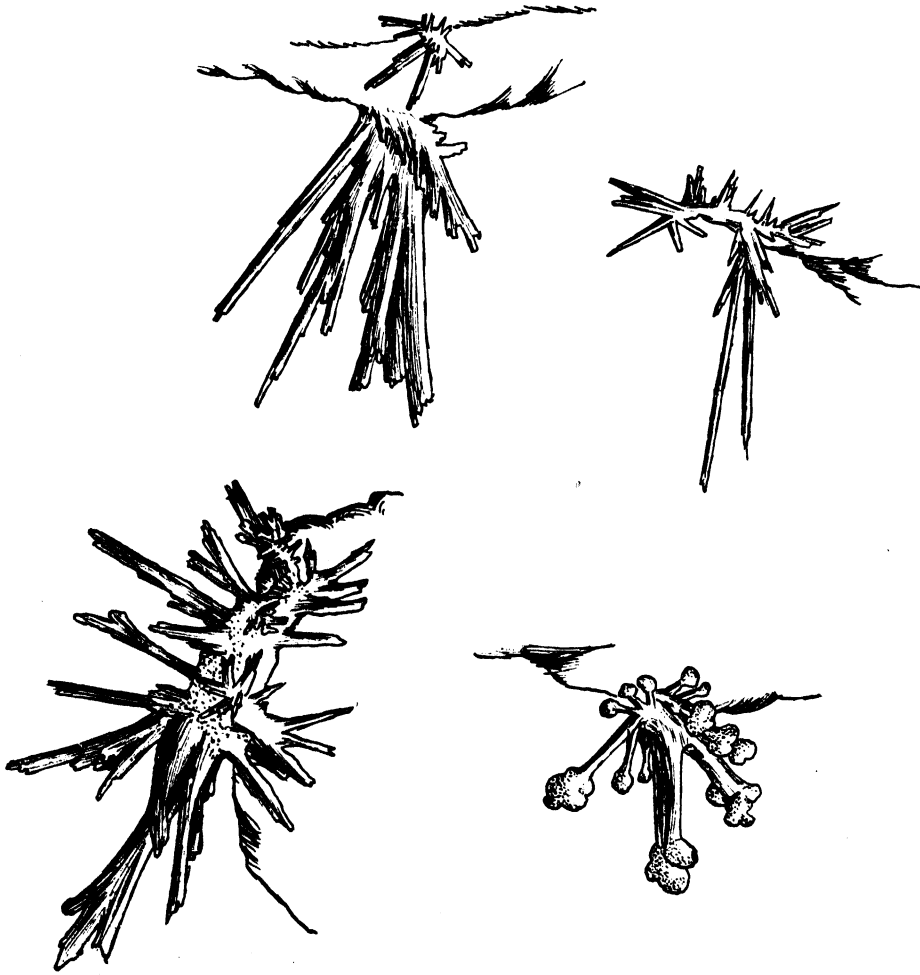

FIG. 3.—Typical helictites (natural size)

Malott[7] postulated another aspect of the Davis theory by asserting that cave systems develop in a spongelike or rambling pattern below the water table. Later these passages are drained, when elevated above the water table, and are then invaded by surface streams which work into the pre-existing cave system and modify the caves according to the more linear nature of their erosion. The Goodwins Ferry Cave is unusually straight and level with no evidence of earlier rambling channels. Further, there is no evidence, on the slope of Spruce Run Mountain, of the remains of any surface stream which could have been located so as to work its way into the cave. As has been indicated, the source of the cave's water is, and probably always has been, from the overlying sandstones. Surface streams are frequently subject to floods, and underground streams of surface origin are also subject to fluctuations in volume due to intermittent rainfall. The lower level of the Goodwins Ferry Cave does not show any evidence of flooding. The ceiling, which in many places slopes down to the stream, is entirely clean and shows no mark of higher water levels. Further, the cave water is clear and, to the author's knowledge, has never been known to become muddy, as would be expected if the water came from a surface stream. It seems certain that the invasion theory in its original form does not apply to the Goodwins Ferry Cave.

Still along a similar line, Swinnerton[8] presented the theory that lateral flow in the upper zone of the water table causes the development of caverns. He, also, was trying to explain originally rambling patterns in caves as well as the horizontal nature of many cave levels. In a limited sense, this theory could apply to the Goodwins Ferry Cave, since the stream in old age would be flowing on the level of a water table. But, since there are no rambling patterns, and the stream's water comes from the overlying sandstones, the lateral-flow theory in its proper form is not applicable to this cave.

The concept of an overlying water table that furnishes the source of a cave stream was advocated by Gardner,[9] who, while admitting other modes of origin, stressed the theory that caves develop from the tapping of static-water zones as surface valleys are deepened in the process of baseleveling. Undoubtedly, the Goodwins Ferry Cave furnishes a path for the drainage of water from the sandstones that make up the top of Spruce Run Mountain down to the New River which determines the local baselevel.

The New River, in cutting across Spruce Run Mountain, has left a steep gorge, on one side of which the entrance to the cave is perched. The extreme levelness of the cave (Fig. 2) indicates that it must have developed when the New River was flowing near the elevation of the cave, which is now about 250 feet above the river level. It has been noted that the cave stream is in the act of lowering itself to the present course of the New River through the lower spring, which drains about half of the cave water and which issues at practically the same elevation as the river. It seems logical to suppose that the New River was at one time stationary at or near the elevation of the cave entrance long enough to furnish time for the cave to develop strongly on

[7] C. A. Malott, "Invasion Theory of Cavern Development," *Proc. Geol. Soc. Amer. 1937*, p. 323.

[8] A. C. Swinnerton, "Origin of Limestone Caverns," *Bull. Geol. Soc. Amer.*, Vol. XLIII (1932), pp. 663–94.

[9] J. H. Gardner, "Origin and Development of Limestone Caverns," *Bull. Geol. Soc. Amer.*, Vol. XLVI (1935), pp. 1255–74.

that horizon and that, since that time, the river has been downcutting more rapidly than the cave stream. However, the author has not been able to recognize any surface indications of a pause in the river's erosion or any terraces or other erosional surfaces near by at elevations that might suggest a geomorphic relation to the cave. It is possible that the cave may be a more delicate criterion of temporary regional baselevel than are surface erosional features. If this is the case, the study of other caves in the region could furnish comparable data on elevations which might prove to be of significant value.

The volume of the cave stream, when it was active in the upper level of the cave, cannot be determined, but, from the present exceedingly steady flow and the source of the water, there is no reason to think that the stream's volume has varied much, barring climatic changes. The hollow through which the stream cascaded when flowing out of the upper level is strikingly similar to the little valley of the present stream farther down the slope, both in size and in talus accumulation. Thus it seems likely that there was a similarity in stream size throughout the time that the cave has existed.

Inside the cave the huge development of the Attic and the other similar enlargement about 150 feet west (Fig. 2) seem incongruous when compared with the remainder of the cave. Excessive rock-falls in the Attic have obscured much of the true nature of that room, but it is evident that much stream erosion took place there at an earlier time. It is highly probable, and in keeping with the rest of the cave, that the cave stream at one time entered the Attic in some manner before the main passage had worked its way beyond the Attic and that the stream flowed from the present tip of the Attic directly across to the previously mentioned enlargement and so developed it. The extension of the Forest Room is probably indicative of another channel, in the same direction, but developed at a later time. Rock-falls and dripstone deposition could easily have sealed off such connections and also obscured evidences of stream entrance into the Attic. Since there are about 200 feet of limestone between the Attic and the source of the stream in the overlying sandstone, it is quite likely that there once existed many ramifications of the cavern that have been lost. The tiny outlet of the Attic into the present hallway indicates that the stream flowed through that opening for only a comparatively short time before it left the Attic entirely and entered the main hallway somewhere beyond its present extremity, which is now blocked by rock-falls. A few abandoned and blocked courses seem to have developed about 200 feet west of the present termination of the main hallway. Further confirmation of the flow from the tip of the Attic across to the hall enlargement is afforded by the excessive height of the main hall west of this enlargement. The height not only is due to flow from two sources but also may indicate a strong slope in the first channel, as there does not seem to be any excessive widening farther west toward the entrance. The development and abandonment of several stream courses in the main level are also shown directly east of the Church Room.

After the mature development of the upper level came the ponding already referred to. For a while it is likely that the stream flowed in the upper level to the Lunch Room, developing that room to its present size, and at the Lunch Room the stream then took the course down to the

lower level. Next in the development of the cave was the channel called the "China Slide," which pirated the stream's water away from the course leading to the Lunch Room and increased the development of the cave's lower level. Following the China Slide cutoff, part of the stream appears to have found access to the lower cave at a distance of more than 300 feet upstream. Here, two moderately sized rooms were developed, and a waterfall of recent origin seems to be tapping a stream in what may be an extension of the upper hallway blocked by debris. A difference in age can be seen on the lower cave level, as there is dripstone downstream from the Lunch Room, but the stream level is devoid of deposits for its entire length upstream. It has already been stated that the stream at present loses much of its water at points downstream from the Lunch Room. Evidently the development of the cave takes place in regularly repetitive steps in which the stream continuously works downward and, once achieving a lower level, backward to regain any length it may have lost.

The waterfall is an extremely youthful feature, judging from its diameter of only about 11 feet. The volume of water crashing down the center of the chimney produces a spray that keeps the cylindrical walls constantly wet, and it is probably this descending film of water that dissolves the wall rock and thus enlarges the chimney. As the chimney becomes enlarged most rapidly downward, the water goes over a lip at the top to fall freely in the center of the chimney. Further development would result in a dome room.[10] Such waterfall tapping of an overhead stream is called "subterranean stream piracy."[11] In this instance water has been diverted from another course into that of a more youthful and more vigorous stream.

The caving of the dolomite roof was an important factor in enlarging the cavern, but the ceiling everywhere appears extremely firm, and it is not apparent why such large rock-falls should have taken place. Of course, enlargement of the cavern may have enabled rock pressures, together with original open fissures in the dolomite, to drop the blocks. But there is a possible tectonic explanation, for the center of a recent major earthquake was only 10 miles from the cave. Following are some excerpts from a report on the shocks:

> Immediately following the earthquake of May 31, 1897, which was distinctly felt over most of the eastern portion of the United States, came reports of continued disturbances in the form of explosions and earth tremors in Giles County, Virginia. . . . . In Pearisburg . . . . no serious damage was done but old brick houses were badly shaken, and many chimneys were cracked and the topmost bricks hurled to the ground. Much noise accompanied the shock. . . . . At Narrows . . . . the surface is said to have rolled like the groundswells of the ocean, springs were muddied and in some cases ceased to flow for a short time after the shock occurred, and a landslide of considerable proportions and a big rock rolled down off the face of Wolf Creek Mountain.[12]

No doubt such shocks could have caused rock-falls in the cave, but whether or not most of the rock-falls depended upon such unusual disturbances is an unanswerable question. Gardner[13] believed that unrecorded earthquakes were important in causing enlargements by caving in Carlsbad Cavern and other west-

[10] J H. Bretz, "Vadose and Phreatic Features of Limestone Caverns," *Jour. Geol.*, Vol. L (1942), p. 683.

[11] *Ibid.*, p. 685.

[12] M. R. Campbell, "Earthquake Shocks in Giles County, Virginia," *Science*, Vol. VII (1898), pp. 233–35.

[13] Pp. 1272 of ftn. 9.

ern caves, but these opinions also cannot be proved.

V. OTHER FAULT CAVES

In the American literature the writer has found no similar caverns described, and, for the present, the Goodwins Ferry Cave and its sister-cavern across the river may be considered as being unique in their association with a major fault.

It has not been possible to investigate the foreign literature thoroughly, owing to the lack of comprehensive foreign bibliographies prior to 1933. The only reference to a similar cave that the writer has been able to find is the report of C. Sorotchinsky on the origin of the Salle du Dôme in the Grotte de Han of Belgium.[14] The origin of the "Dome Room" is attributed to a combination of faulting, caving, and solution, but it differs from the Goodwins Ferry Cave in that the room does not lie on the fault itself but is on one side of a faulted zone. Enlargement of the room has resulted from caving of the ceiling in a limestone bent into a monocline. The Dome Room has a height of over 100 feet and is located under the crest of the monocline which made the enlargement possible. The room is not associated with the fault, but near by is a stream in a constricted passageway ("passage du Diable") which appears to be well in the fault zone. The fault cannot be defined as a plane but is represented by a zone of brecciation.

[14] C. Sorotchinsky, "Un accident téctonic éclairant la génèse de la Salle du Dôme dans la Grotte de Han," *Bull. Soc. Sci. Bruxelles*, Vol. LIX (1939), pp. 97–106.

CONCLUSIONS

The Goodwins Ferry Cave is the result of solution along the plane of weakness of the Saltville fault. The cave stream has worked its way down the dip of the fault and along the fault strike, chiefly in the limestone. Rock-falls from the roof have enlarged the cavern into the dolomite, and, as the cavern works its way down the dip, dripstone is acting to fill in the cavern from above. The linear shape of the cave is due to its location on the fault plane, and the strongly horizontal development may be due to an earlier control of the cave's drainage by the New River.

ACKNOWLEDGMENTS.—The author is indebted to Dr. J. W. Huddle, of the University of North Carolina, for much valuable assistance in the preparation of this paper. Dr. B. N. Cooper, of Virginia Polytechnic Institute, has given advice on the field work; and field assistance was received from Messrs. S. C. Rainey and I. R. Tannenbaum of the Pioneer's and Hobo's Union. The writer also deeply appreciates the hospitality of Mr. A. B. Scott of Goodwins Ferry, Virginia.

REVIEWS

Geology of Florida. By C. WYTHE COOKE. (Florida Geological Survey Bull. 29.) Prepared in co-operation with U.S. Geological Survey. Tallahassee, Fla., 1945. Pp. ix+339; pls. 1 (map in pocket, scale 1/1,000,000); figs. 47.

It has been sixteen years since Cooke and Mossom's *Geology of Florida*, published as *Florida Geological Survey Twentieth Annual Report*, first appeared. During that time a large amount of new information has been gathered dealing with the geology of Florida; and, although much of it has been published elsewhere, there was great need for a new bulletin bringing this information together in one place and interpreting it in the light of the latest knowledge.

Dr. Cooke has done this in Bulletin 29, which bears the same title as his and Mossom's earlier report.

The new bulletin contains much that is recognizable in the older report with only such references to it as Cooke believed to be of historical interest. As he remarks in the Preface, "It contains many paraphrases and verbatim excerpts from that report without quotation marks or other indications of origin." Thus, Bulletin 29 is more a revision of an old, standard work than an entirely new treatise. It is essentially a stratigraphic paper; however, its contents are not limited to stratigraphy. Five pages are given to a lucid description of the composition and structure of the Floridian Plateau, six pages are devoted to topography, and four pages describe the effect of an oscillating sea-level during Pleistocene time. The Bibliography requires eleven pages and the Index twelve; the balance, two hundred and ninety-seven pages, describes the stratigraphy.

Cooke describes each formation according to the following plan: A, general features: (1) name, (2) characters, (3) thickness, (4) distribution, (5) stratigraphic relations, (6) paleogeography, (7) fauna, and, occasionally, (8) utilization; and B, local details (which includes formation descriptions by counties). In general, the plan and treatment of the subject are excellent.

Since the writing of the *Twentieth Annual Report* a considerable amount of exploration of the subsurface formations by means of test-well drilling has been done by oil companies and independent operators over much of the state in a search for petroleum and by the U.S. Geological Survey in southern Florida on ground-water research. The data uncovered have led to recognition of several new stratigraphic units and, together with new discoveries of surficial relationships, have aided greatly in the solving of several problems in stratigraphy and in the geological history of the state. These new findings are ably discussed by Cooke.

Owing to a lack of printing funds, the geological map of the state (Pl. I) as a whole does not compare favorably with the earlier geological map published to accompany the *Twentieth Annual Report*. The scale would have been more acceptable, especially to petroleum geologists, had it been the same as that used by the state geological surveys of Alabama and Georgia. Use of colors in patterns of blue and orange is not as satisfactory as the color scheme used on the old map, nor is the means of showing the distribution of Pleistocene formations entirely suitable. For example, the Anastasia formation, which is largely contemporaneous with the Miami oölite and Key Largo limestone, is included with the higher terrace deposits under the label "Late Pleistocene Deposits." To be consistent, the Miami oölite and Key Largo limestone also should have been included. In addition, there are vast areas of the state that are covered with a mantle of Late Pleistocene materials; these are now shown but could have been as easily as the areas delineated. It would seem to have been better either to have eliminated entirely the higher Pleistocene terrace deposits (as Cooke did for much of the state) or to have shown their presence over the underlying formations by use of a distinctive overprint.

The route of part of the proposed Florida ship canal is indicated prominently on the map; however, the canal is not yet constructed nor is there any assurance that it will be built. The route of the proposed canal as shown on the map may also lead one to wonder if the eastern terminus is to be at St. Augustine or at the mouth of the St. John's River.

In general, Bulletin 29 is an excellent summary of the stratigraphy of Florida as shown at the time of writing. It fills a need felt by professional geologists, amateurs, and laymen since the *Twentieth Annual Report* went out of print several years ago.

GARALD G. PARKER

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THE VREDEFORT RING-STRUCTURE OF SOUTH AFRICA

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ABSTRACT

Five theoretical explanations of the marvelous deformation in the Vredefort region, all ascribing it to purely terrestrial forces, are summarized. Two of the hypotheses assume magmatic upthrust to have been the primary cause. A third finds it in tangential force, of the kind represented in orogeny of the ordinary type. The remaining two regard horizontally directed force and magmatic upthrust as both essential, though with differing emphasis on relative importance. These wide contrasts among the five interpretations of the unique "ring-structure" illustrate, of themselves, the difficulty of finding explanation in terms of orthodox geological dynamics. It seems, therefore, advisable to take seriously the published suggestion of Boon and Albritton that the structure may have been developed by the infall of a speedy meteorite of planetoid or asteroid dimensions, followed by readjustments of the scarred and shattered terrain that were induced by the earth's gravitation. The effort to compare reasonable deductions from that double premise with the known facts of the field is described in the second half of the paper.

INTRODUCTION

South Africa is crowded with geological wonders. Not the least startling is the "ring-structure" of the "Vredefort mountain land," crossed by the boundary between the Transvaal and the Orange Free State. Concerning its origin, widely contrasted hypotheses have been published during the last forty years. The last to appear in print is that of B. D. Maree.[1] With the gravimeter he has attempted to complete the map of this structure if it were more fully exposed—that is, after the removal of Karroo sediments which still unconformably cover about one-third of the structure. Studying Maree's report, the present writer, who in 1922 had traversed the area under the expert guidance of A. L. Hall of the Union Geological Survey and of the late G. A. F. Molengraaff, formerly chief geologist of the Transvaal Republic, was led to review also the older views of origin. The gamut of all these opinions is wide and of itself strongly suggests how hard it is to understand the Vredefort structure by appeal to recognized terrestrial forces. Perhaps because of the writer's recent concern with the problem of the lunar "craters," there flashed a question: Does the Vredefort ring-structure represent the deeply eroded remains of a meteorite scar? Soon memory was stirred; it was recalled that J. D. Boon and C. C. Albritton, Jr.,[2] had asked the same question in print, though they did not venture an adequate discussion of it. With the double purpose of advertising the

[1] "The Vredefort Structure as Revealed by a Gravimetric Survey," *Trans. Geol. Soc. So. Africa*, Vol. XLVII (1944), p. 183.

[2] "Meteorite Scars in Ancient Rocks," *Field and Laboratory*, Vol. V (1937), p. 53.

Linlithgow Library.

Boon-Albritton idea and also the absence of general satisfaction with the appeals to terrestrial forces as dominant in the structure's development, the following pages have been written. First, the peerless ring-structure itself will be described. There will follow a brief account of the other published views about origin. Finally, the meteorite-scar idea will be somewhat elaborated, in the hope that it may be further examined by those more competent to judge the effects of the infall of a giant bolide (meteorite, asteroid, or planetoid) on the pre-Karroo terrain of the Vredefort region.

THE RING-STRUCTURE

The rocks essentially constituting the Vredefort structure consist of: (1) a thick packet of unfossiliferous quartzites, shales, conglomerates, and interbedded basic lavas—all generally assigned to the late Pre-Cambrian; and (2) "Old Granite" (enclosing some masses of schists), underlying the packet unconformably and now assigned to the early Pre-Cambrian. At some epoch long before the Carboniferous, perhaps as early as the late Pre-Cambrian itself, the amazing deformation of the packet and its granite floor took place. How ever formed, the new structure must have had considerable relief and ruggedness. It was subjected to erosion, more or less continuously, and thus peneplained. In Upper Carboniferous time this peneplain was buried under a relatively thin cover of Karroo sediments, which still remain flat, little warped out of their original attitude. In post-Jurassic time the Karroo cover has been stripped off about two-thirds of the Vredefort ring-structure.

L. T. Nel[3] has made the most complete study of the stratigraphy. From his report and from that of A. L. Hall and G. A. F. Molengraaff[4] Table 1 has been compiled, showing the succession and thicknesses in the stratified mass of the packet, the Pretoria series being at the top.

The stratified beds are seen now to stand at high angles with the horizontal plane, the outcropping edges of the whole "packet" forming a "collar" wrapped about a "core" of Old Granite (see Fig. 1). As exposed, the unconformable contact of core and collar approaches that of a circle 40 kilometers in diameter.

TABLE 1

SUCCESSION AND MEAN THICKNESSES OF THE BEDDED ROCKS

(In Meters)

Pretoria series	3,600
Dolomite series	1,525
Ventersdorp system (basic lavas)	3,700
Upper Witwatersrand division (quartzite, shale [slate], and conglomerate, with diabase sills)	3,900
Lower Witwatersrand division (quartzite, shale [slate], lava beds, with diabase sill)	3,500
Total	16,225

However, according to Maree's gravimetric results (apparently confirmed by data of D. J. Simpson using the magnetometric method of detection)—see L. T. Nel[5]—that contact on the southeast bends out of the circle in such a way that the ground-plan of the core surface is like that of the flat section of a pear which has been sliced longitudinally. Accordingly, Maree puts the

[3] *The Geology of the Country around Vredefort: An Explanation of the Geological Map* (Pretoria: Geological Survey of South Africa, 1927).

[4] *The Vredefort Mountain Land in the Southern Transvaal and the Northern Orange Free State* (separately printed in "Shaler Memorial Series," from *Verhandl. Kon. Akad. Wetenschappen Amsterdam*, Sectie 2, Deel 2 [1928]).

[5] "Remarks in Discussion of Paper by D. W. Bishopp," *Trans. Geol. Soc. So. Africa*, Vol. XLIV (1941), p. xciii.

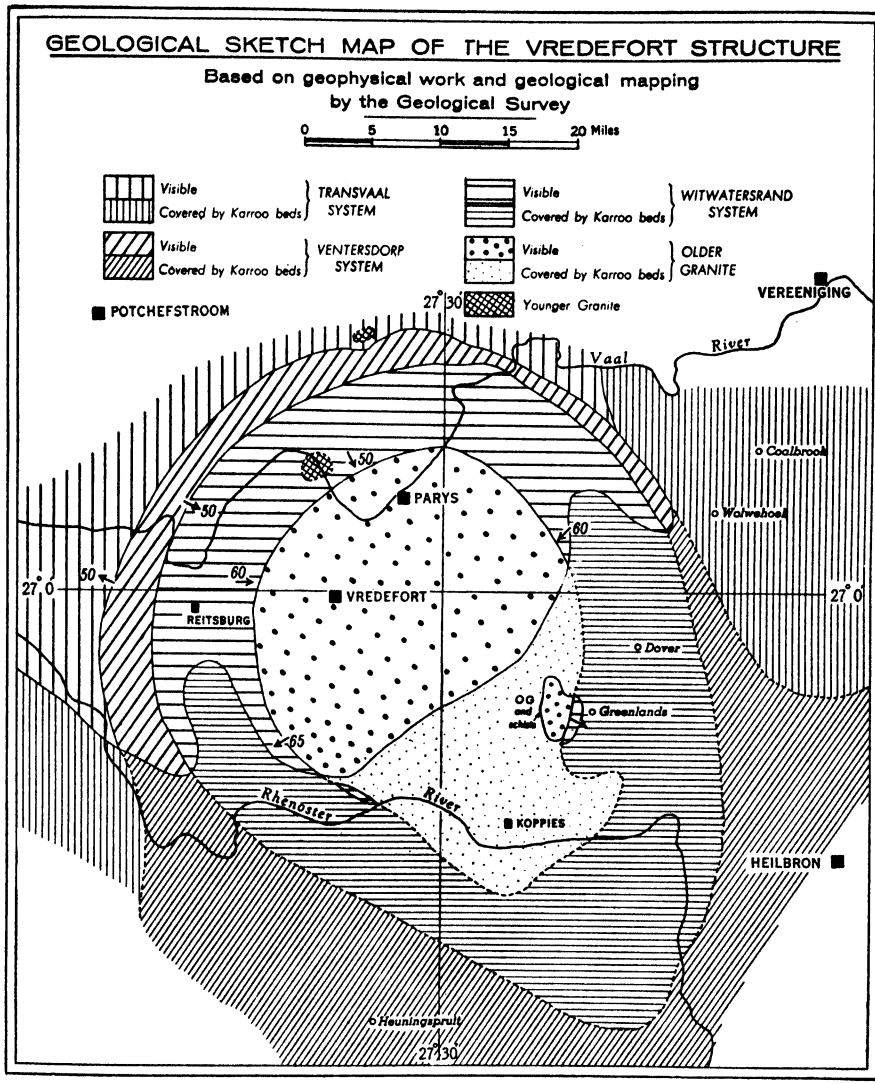

FIG. 1.—Generalized map of the Vredefort region, drawn by L. T. Nel (see ftn. 5) to illustrate the ring-structure and its southeastern extension under the remnant of the Karroo cover, as indicated by the gravimetric method of detection. (Copied by permission of the Geological Society of South Africa.)

maximum, northwest-southeast, diameter of the granitic surface, supposed freed from the Karroo cover, at 50 kilometers.

Along lines radiating out from the center of the granitic area, the collar has outcrop widths ranging from 15 to 20 kilometers.

Nel's large map indicates many observed dips and strikes read within the Witwatersrand divisions. The angles of these dips have been averaged in the three quadrants of the visible structure, mutually separated by the meridian of 27°30′ east longitude and the 27° parallel of south latitude. The averages are shown in Table 2, where the number of readings for each average is given in brackets. Besides vertical dips, those directed in a general way toward the center of the granite core and also those directed in the opposite general direction are separately tabulated.

TABLE 2

SUMMARY OF DIPS (AVERAGES)

	Vertical (1)	Inward (2)	Outward (3)	Grand Average (All Inward) (4)
Northwest quadrant:				
Lower Witwatersrand	[5]	[90] 53°	[8] 60°	[103] 61°
Upper Witwatersrand	.....	[61] 50	[4] 57	[65] 54
Northeast quadrant:				
Lower Witwatersrand	[1]	[33] 55	[4] 65	[38] 62
Upper Witwatersrand	.....	[28] 56	[1] 60	[29] 58
Southwest quadrant:				
Lower Witwatersrand	[1]	[21] 57	[13] 66	[35] 79
Upper Witwatersrand	[1]	[7] 66	[7] 70	[15] 88

The statistics of Table 2 show that in the northerly quadrants the whole of the Witwatersrand system has not only been upturned but also strongly overtilted. The South African geologists seem to agree that the same is the case with the Ventersdorp system. In the Dolomite series on the north, Hall and Molengraaff observed four "inward" dips and regarded them as registering overtilt even at this great distance from the granite core, though those authors gave no direct evidence of such violence of deformation for the Dolomite itself.

From his gravimetric results Maree[6] concludes that, in general, overtilting characterizes the strata only at and near the surface, the dips becoming "outward" at relatively small depth. He does, however, recognize strong overtilting to depth in the northwest quadrant of the ring-structure (see the sections of Fig. 2).

Clearly the stresses set up during this energetic "diastrophism" were enormous, and it is not surprising that Hall and Molengraaff found all the rocks

> to show more or less a cataclastic structure and to be cracked and crushed in many places. The effects of pressure are as a rule concentrated along crush-zones in which they get so strong that the rocks are mylonized or triturated..... The crushing and grinding of the rocks were accompanied by comparatively little [mass-] movement, and although the rocks affected are very rich in cracks and micro-faults, schistose structures are rarely found.[7]

It was to be expected that detailed mapping of the collar would display

[6] P. 191 of ftn. 1.

[7] P. 148 of ftn. 4.

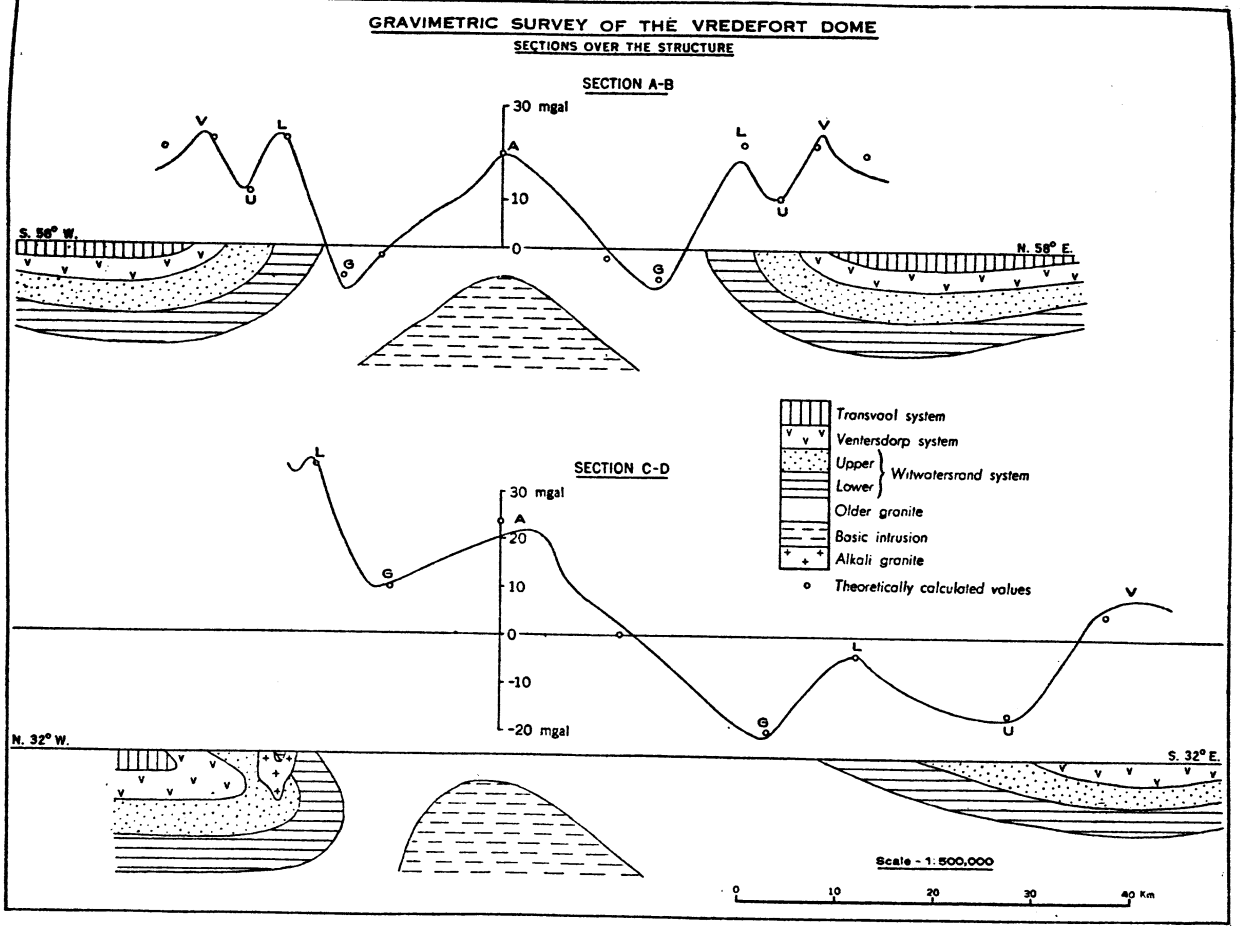

FIG. 2.—Sections from Figure 3 of B. D. Maree's paper, ftn. 1. The lettered curves show the run of gravity anomalies. Those at lettered points were found over outcrops or inferred positions of edges of the various rock systems under the Karroo cover, as follows: *V*—on the Ventersdorp system; *U*—on the Upper Witwatersrand division; *L*—on the Lower Witwatersrand division; *G*—on the Old Granite (core). *A* represents the anomaly above the top of the inferred basic intrusion.

The structure sections, at right angles to each other and with indicated azimuths, are diagrammatic and drawn to illustrate the position of

much macrofaulting; in fact, Nel's map shows many hundreds of observed faults, and there must be many more in belts poor in outcrops. A host of similar dislocations must be represented in depth. The observed faults run in all azimuths on the map.

The intensity of the deformation is further suggested by some of the rapid changes of measured thicknesses of each member of the two Witwatersrand divisions of the collar strata, as these are traced along the curved outcrop. Nel's tables[8] give the ranges of thicknesses. While most of the ranges are of the order expected, because the formations in this thick geosynclinal prism were lenticular from the start, some are not to be so readily explained.

TABLE 3

ILLUSTRATING VARIATION OF THICKNESS

Locality (Farm)	Thickness (In Feet)	Approximate Distance Apart (In Miles)
Procedeerfontein No. 110...	6,000	
		4
Vaalwal No. 158.....	7,800	
		14
Koedoesfontein No. 12, etc..	1,300	
		7
Rooderand No. 26, etc.....	9,000	
		4
Nooitgedacht No. 56.....	9,700	
		3
Elandslaagte No. 28.....	9,500	

Concerning the Orange Grove series, Nel wrote:

> Great pressures acting on the group of quartzites and slates have pinched out one or more of the latter in certain localities..... [On account of] the great pressures to which the whole system was subjected in the course of being overtilted and faulted, and so yielded to the stresses set up, ... [the Jeppestown] series in some localities became squeezed out to such an extent that the total thickness may be only 800 feet, instead of the 1,600 feet calculated for the average thickness. This, apart from any lenticular habit the strata may possess, probably accounts for the considerable discrepancies in the thickness of the series here.[9]

A second example is to be suspected in the great differences of thicknesses stated for a composite quartzite-slate member of the Kimberley-Elsburg series (see Table 3).

Such squeezing-out would have been a probable accompaniment of the inevitable stretching of the strata as these were upturned and overtilted.

Incidentally, it may be asked whether the total thickness of the collar strata before their deformation was not greater than the total measured at the existing outcrops.

In spite of all the complications, the average trends of the strikes, like the upstanding ribs of hard quartzite, are eloquent of the fact that the upturning and the overtilting of the beds were controlled by forces directed along lines radiating from a central or subcentral sector of the Old Granite core.

Either after or during the deformation of the strata the collar was invaded by magmatic bodies, whose order of emplacement is given by Nel[10] as follows, beginning with the oldest: (1) basic intrusions (sheets and dikes of gabbro and diabase with subordinate pyroxenite, hyperite, and peridotite); (2) "younger" alkaline-granite stocks; (3) nepheline-syenite dikes and veins; and (4) enstatite-granophyre dikes. Large dikes and sheets of gabbroid rock were injected also into the Old Granite of the core, particularly within a relatively narrow belt along the contact of core and collar.

The younger, alkali-rich, granite occurs in the form of three stocks. Their

[8] Pp. 33, 35, 39, 40, 43, 46 of ftn. 3.

[9] P. 33 of ftn. 3.

[10] P. 60 of ftn. 3.

contacts cut across the strata of the collar in its northern part (see Fig. 3), but at least in one case the Witwatersrand beds seem to have been bulged up so as to form "an irregularly shaped dome or anticline."[11] The nepheline syenites are closely associated in chemistry and time of intrusion with the younger granite. The enstatite-granophyre dikes are highly inclined and cut across the contact between core and collar as well as across the Old Granite well within that contact. Nel agrees with Hall and Molengraaff in taking the view that these peculiar dikes are large-scale injections of the once-vitreous material like, or analogous with, that represented in the countless veins and small irregular masses of pseudo-tachylyte, which S. J. Shand[12] has shown to be local re-melts of the Old Granite.

[11] P. 68 of ftn. 3.

[12] "The Pseudotachylyte of Parijs," *Quart. Jour. Geol. Soc. London*, Vol. LXII (1917), p. 198.

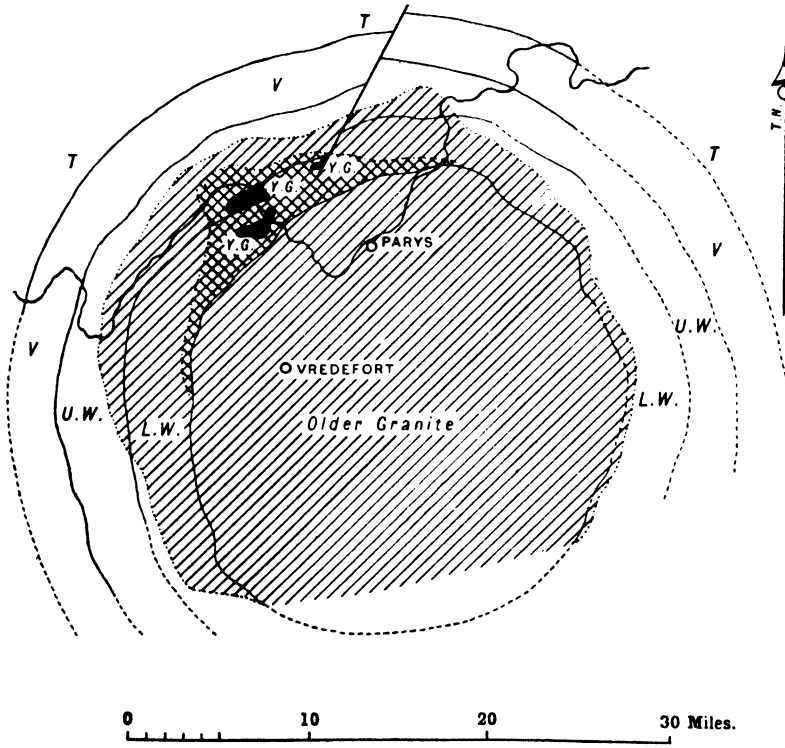

FIG. 3.—L. T. Nel's (see ftn. 3) diagrammatic map of the ring-structure. (Copied by permission of the Geological Society of South Africa.) *T*—Transvaal system; *V*—Ventersdorp system; *U.W.*—Upper Witwatersrand division; *L.W.*—Lower Witwatersrand division; *Y.G.*—younger granite.

The area covered by single-line shading denotes the distribution of metamorphosed rocks and Old Granite, while the portion crosshatched denotes the area of most intensely altered sedimentary rocks.

Broken lines indicate extension of formations under covering of coal measures (Karroo).

The completion of the circle on the south by the broken line is entirely hypothetical, there being no evidence from boreholes that the boundary of the granite is as shown. (The geophysical data of Maree and Simpson were not available in 1927, the year when Nel published this diagram.)

One more principal fact about the collar remains to be mentioned, namely, the advanced thermal metamorphism of the shales and amygdaloidal lavas occurring in the Lower Witwatersrand division and, in the northwest quadrant of the ring-structure, extending into the Upper Witwatersrand division (see Fig. 3). In that region is the locus of the three bodies of younger granite. There, however, the width of the metamorphosed belt is out of all proportion with that expected from the heat and gases of this granite, unless the eruptives coalesce just below the present surface to form a body very much more extensive than all three of the exposed bodies put together. Nel[13] has drawn that inference and actually credits the injection of that greater, largely invisible mass of young-granite magma with being a leading factor in the main deformation at Vredefort.

In its turn the core of Old Granite gives evidence of the energy that must have been expended when the ring-structure was developed. As Hall and Molengraaff noted, this granite "has been subjected to considerable pressure, as shown under the microscope by cataclastic structure. Not seldom the crystals of quartz and feldspar show many cracks filled with a mosaic of crushed material. Nowhere, however, has the pressure given rise to a distinct schistosity."[14] Doubtless connected with that intimate crushing is the widespread macroscopic shattering of the granite, the shatter-blocks being separated or enclosed by the veinlike masses of pseudo-tachylyte. Shand's conclusion that the latter was formed by local re-melting seems to be generally acceptable. The exceedingly fine grain of the pseudo-tachylyte is well explained by A. Holmes[15] as due to the low content of volatile fluxes left in the Old Granite after its own, earlier, crystallization. Assuming that judgment to be sound, one can hardly avoid the deduction of about 700° C. as a minimum temperature for the Old Granite when the partial melting took place. It is clear that the required temperature was not in any large measure the result of shearing of the granite. Heating by pure crustal pressure without plastic flow has been suggested as the cause of the extra heating, but that seems to be a quite inadmissible explanation. Shand left the problem open, though narrowed down to two alternatives. He wrote: "The pseudotachylyte has originated from the granite itself through melting, caused (as I have shown) not by shearing but by shock, or, alternatively, by gas-fluxing."[16]

FIVE THEORIES OF ORIGIN

Five pictures have been painted in print to account for the ring-structure through the operation of purely terrestrial forces. As already remarked, the five hypotheses disagree with one another, and such disagreement among serious and competent workers creates doubt that reliance on the orthodox, textbook principles of geology alone is justified. To illustrate the point best, the five hypothetical pictures will be examined individually.

1. In their book Hall and Molengraaff[17] outlined the various explanations of the ring-structure that were published before 1922. Because those skilled observers made the first detailed map, their theory is the first to deserve attention. They attributed the mighty deformation to two kinds of pressure. The dominant

[13] P. 90 of ftn. 3.

[14] P. 17 of ftn. 4.

[15] See discussion on p. 221 of ftn. 12.

[16] P. 219 of ftn. 12.

[17] See ftn. 4.

kind was horizontal, affecting at least the outer part of the earth's crust as a whole or else over a wide stretch. The other kind of pressure, subordinate but still important, was that causing the vertical upthrust of the core of Old Granite. At the opening of their theoretical argument the two authors presented their vision of what had happened in the Vredefort area as follows:

In the doubtless very great lapse of time which stretches between the end of the deposition of the sediments of the Transvaal System and the beginning of the deposition of sediments belonging to the Karroo System a peculiar, well localized, and very powerful diastrophism took place in the Vredefort Mountain Land. Consequent on these crust-movements all round a common centre now occupied by the Vredefort [Old] Granite, all the sediments ranging from the base of the Witwatersrand System up to the top of the Transvaal System were tilted and largely overtilted. In the centre of the disturbed area the Vredefort Granite was uptruded to a level which, before it was lowered again by denudation, certainly must have been much higher than that occupied by the uppermost sediments of the Transvaal System. The result, as it can be seen now, is a boss of granite, the Vredefort granite-boss encircled by a girdle of tilted and overtilted sediments. The total vertical displacement of the granite in the central portion of the Vredefort dome may be estimated as more than 14,000 m.

The granite was uptruded at the same time that the strata of the sediments were tilted and finally overtilted. What is the relation between these two phenomena? Has the granite played an active role in the process of updoming, or has it been moved upwards passively, following the uplift of the sedimentary roof?[18]

The hypothesis that the updoming was caused primarily by the upward push of a large body of deep-seated magma on the plug or core of Old Granite was considered and rejected as quite inadequate. Less mental trouble is encountered if one assumes that the primary cause of the updoming was an orogenetic one and that centripetal stresses in the earth's crust caused the horizontal strata and with them the underlying floor of granite and schists to be elevated and arched up into a huge circular dome. If one admits the plane of application of the centripetal forces to have been lower than the present visible surface of the granite, an overtilted position of the strata in the lowermost portion of the slope of the dome could be expected. Further, the intense stress, the effects of which are visible in all the rocks throughout the Vredefort area, is, on this assumption, of necessity the result of the powerful pressure which resulted in the updoming of a huge block of the earth's crust.

The magma in the depth below the central granite was released by the local relief of pressure due to the updoming, and got the opportunity to invade with force the much heated and probably semiplastic granite floor as well as the roof of the dome. Differences in the capacity to which the various roof strata were able to resist pressure would give the ascending magma during the updoming an opportunity to be injected as sills.[19]

On pages 162–63 of their book is a summary of the final hypothesis put forward by Hall and Molengraaff:

The present structure of the Vredefort Mountain Land is based on centripetal pressure as the major cause, associated with a "point" uplift as a minor cause and dependent upon the emplacement of a younger magma. The combined effect is the updoming of the central granite as a more or less passive body. . . . . As regards the time order, the authors consider that the centripetal pressure came into force first, to be followed later on by movements of magma.

Hall and Molengraaff knew well the toughness of the genetic problem. They pointed out that "the initial cause of the centripetal pressure in the earth's crust remains unexplained."[20] They could find no good reason why the ring-structure should have been developed in the midst of a great geosynclinal prism of sediments. Although their cross section of the

[18] Pp. 144–45 of ftn. 4.

[19] P. 155 of ftn. 4.

[20] P. 163 of ftn. 4.

ring-structure (Fig. 1 of their book) is diagrammatic, it seems to indicate a central uplift of at least 30 kilometers and therefore a departure from isostatic balance very much greater than any yet demonstrated in the existing earth.

In view of all these deficiencies the Hall-Molengraaff hypothesis can hardly be accepted as a sound explanation for the extraordinary features of the ring-structure.

2. Toward the end of his book Nel agreed that the nature of the forces that resulted in the "updoming of the rock formations" is "imperfectly understood." He wrote:

> The Vredefort structure may be due either to magmatic upheaval or to centripetal pressure. . . . . Future investigations and research may throw more light on this subject. Data may be collected which will indicate which of the two, horizontal compression or magmatic upheaval, was the chief agent which actually produced the updoming. At present it appears that the two were really closely associated.[21]

He had a degree of skepticism about the idea of centripetal movements of the rocks under converging pressures. On page 109 we read:

> Now tangential forces generally tend to produce elongated upheavals or anticlinal structures, and it may be difficult to picture, therefore, just how a round and practically symmetrical structure like the Vredefort occurrence could be formed, unless the horizontal compression was very evenly distributed. The existence of evenly distributed radial pressures, however, would be most unusual.

On the other hand, Nel emphasizes magmatic upheaval and suggests that the magma concerned was that of the younger alkali-rich granite, extended underground as a "batholith" which was about as wide as the visible surface of the Old Granite core. He assumes that the upthrust would cause the pronounced overturning, eversion, of the sedimentary beds overlying the Old Granite.

The reasons for making this last assumption need clarification. Furthermore, the validity of the magmatic hypothesis must be questioned until it be coupled with a satisfactory theory of origin for the "batholith" itself and for the assumption that it would have had the urge to rise so high into the earth's crust in spite of the great weight of Old Granite and its sedimentary cover. In other words, Nel did not succeed in removing the objections against the double "postulate" of Hall and Molengraaff.

3. B. D. Maree[22] reverses the emphasis and thinks that the deformation was due to centripetal, "tangential," pressure of the kind referred to in theories of ordinary orogeny, magmatic invasion of the earth's crust being unessential to the upturning and overtilting of the sedimentary beds. He makes a valiant attempt to explain the assumed convergence of the "tangential" pressures toward the center of the ring-structure, but the corresponding argument is so charged with uncertain assumptions that the result is far from convincing. The new gravimetric evidence that the ring-structure is pear-shaped, rather than circular, in ground-plan seems overstressed, and Maree's suggestion that successive pressures, applied at two different assumed epochs, would have sufficient centripetal quality appears doubtful.

Maree accounts for the magmatic intrusions into the ring-structure (supposed complete before this invasion) by appeal to "isostatic inequilibrium," induced by the major deformation. He also leaves unsolved this problem of the origin of the magma, which he assumes "because of isostatic considerations" to have "segre-

[21] P. 108 of ftn. 3.

[22] See ftn. 1.

gated into basic and acid phases." Few petrologists would be willing to guess how such segregation could have been accomplished under the given conditions. Even if it did take place, it is hard to follow the reasoning embodied in the following passage in Maree's paper:

> It is noteworthy for isostatic considerations that the magma segregated into basic and acid phases and that the lighter acid alkali-granite was injected into the heavy sediments where there was a relative surplus of mass, and the heavy basic gabbro was injected into the light Older granite where there was a relative deficiency of mass. . . . . Thus, the isostatic equilibrium was adjusted by the segregation and migration through replacement of material, with the result that the whole Vredefort structure and the [surrounding] geosynclinal basin are to-day in relatively good isostatic equilibrium.[23]

4. In complete opposition to Maree's opinion, E. B. Bailey[24] concluded that "a hypothesis of magmatic upheaval and concomitant centrifugal pressure seems much more applicable at Vredefort than one of centripetal pressure."[25] After referring to the difficulty which Hall and Molengraaff felt in accounting for the overtilt of the collar strata, Bailey remarked:

> It may, however, be pointed out that it is impossible to communicate an upward push through a plastic column without at the same time developing a centrifugal push. If the upward push raises the overlying rocks, the outward push (at any fixed point) will steadily increase. It will also produce outward deformation, if this deformation can be accommodated by being indirectly communicated upwards so as to find relief in the raising of the roof. Subterranean outward deformation will laterally extend the superficial dome and ameliorate its slopes.[26]

Bailey attributes the upthrust at Vredefort to the "younger magma" of Hall and Molengraaff.

To support his view of the matter, Bailey drew a parallel with the deformation of schists and sedimentary beds by upthrust of the Arran granite boss in Scotland, but in the section illustrating the Arran "parallel" none of the beds is shown as overtilted and dipping toward the intruded granite. Corroboration for Bailey's idea about Vredefort is not obvious from either the section or the accompanying text.

If it be assumed that the "younger magma" under the Vredefort area pushed up the Old Granite and its sedimentary cover to form a dome with central height of, say, 15 kilometers, it would be conceivable that the collapse of the dome under its own weight might have given the strata "inward" dips, but it seems incredible that the angles of these dips would be as low as those observed at the existing surface of erosion. Then, too, Bailey's hypothesis can hardly be generally accepted until some reasons are adduced for the localization of the magmatic invasion and for the enormous upthrusting power assumed for it. Neither of those reasons is given or to be readily suggested by even prolonged consideration of the problem.

5. D. W. Bishopp[27] also rejected the postulate of centripetally directed pressure as playing an important role and tried to find the essential solution in magmatic upthrust of the core of Old Granite. He wrote:

> The dynamics of the structure appear to be fairly simple in general principle. A minor uplift in the bottom of the primary shallow basin [the broad geosynclinal prism of which our "collar" sediments formed a part filled this

[23] P. 196 of ftn. 1.

[24] "Domes in Scotland and South Africa: Arran and Vredefort," *Geol. Mag.*, Vol. LXIII (1926), p. 481.

[25] P. 484 of ftn. 24.

[26] P. 483 of ftn. 24.

[27] "Geodynamics of the Vredefort Dome," *Trans. Geol. Soc. So. Africa*, Vol. XLIV (1941), p. 1.

basin] seems to have been associated with high temperatures from a local underlying magma. . . . . An upward pressure from this magma, which need not have been excessively high, first placed the apex of the minor uplift under tension, and caused its rupture with the extrusion of plasticized material into and through it. The process continued, and the sediments were still further upturned and thrust apart.

The upward pressure of the magma on the lower portion of the basin now brought another set of forces into play. The upward swing of the inclined and rather rigid sediments forming the flanks of the structure created a toggle-action, and induced immense longitudinal compression in them, many times greater than the thrust of the magma generating it. As a result the fractured ends of the sediments were still further uplifted and overturned, and the plastic Old granite between them was constricted and squeezed, its uplift being further intensified thereby. There the process ended, from the structural point of view.[28]

Like Bailey, Bishopp does not explain why the magmatic body was developed and located under the Vredefort area or why it had the assumed power of upthrust. The published argument is weakened by these gaps in reasoning, which the present writer finds to involve an apparent inconsistency. Bishopp assumes the Old Granite plug to have been plasticized,[29] and yet, when invoking "toggle-action" as the condition for the centrifugal pressure supposed to cause the strong upturning and overtilting of the sedimentary beds, he assumes a high degree of strength for the same granite when under much greater (horizontal and centrifugal) stress. Further, it is not easy to see how the toggle principle can apply at all after the bared top of the granite plug has reached a horizontal diameter exceeding 5 or 10 kilometers.

The foregoing sketch of five hypotheses erected to account for the Vredefort ring-structure proves the great diversity of opinions about the nature of the deforming forces. The corresponding kinds of pressure may be summarily described as in Table 4.

Hall and Molengraaff and also Nel leave open the question as to the cause of the horizontal pressure. All three realized the difficulty of imagining how such pressure could become centripetal. Maree referred it definitely to the crustal condition assumed in the ruling theory of orogeny—this kind of pressure being here with some arbitrariness called "tangential." His success in showing how it became centripetal with reference to the Vredefort granite core has been found highly doubtful.

With the exception of Maree, all the authors assume as vital the upward magmatic pressure on the visible core of Old Granite, but none of them suggests a cause for the development of the big magmatic body under the Vredefort area or a reason for the assumed gigantic displacing energy of the body. There can be no doubt about the upthrust of the Old Granite core, for at the existing surface of erosion its top is on a level with the Pretoria series of beds outside the ring-structure; but to suppose that an underlying magmatic mass had within itself a great upthrusting power is manifestly hard to believe. On the other hand, the meteorite-scar hypothesis, soon to be described, seems competent to account for the localization of the upthrusting, deep-lying magma on a big scale, this magma having been pushed up, passively, during an isostatic adjustment of pressures still farther down in the earth. The meteorite-scar hypothesis, like those of Bailey and Bishopp, emphasizes centrifugal pressure as the cause of overtilting of strata but is not troubled by the necessity of proving that the magmatic upthrusting pressure could be resolved into strong, centrif-

[28] P. 16 of ftn. 27.

[29] Pp. 7, 10, and 15 of ftn. 27.

ugal, horizontally directed pressure as is the case with each of the Bailey and Bishopp hypotheses.

Hall, Molengraaff, Nel, Maree, and Bailey postulate updoming on a scale unmatched in any other known crustal structure, so far as the height of the dome is concerned. Bishopp postulates upturning without doming in anything like the same amplitude. It will be seen that the meteorite-scar hypothesis postulates deep basining at the surface, with concomitant upturning and ultimate overtilting of the sedimentary beds of the collar.

TABLE 4

Authors	Kinds of Pressure Assumed	
Hall and Molengraaff	Horizontal and centripetal	Magmatic and vertical
Nel	Horizontal and centripetal(?)	Magmatic and vertical
Maree	"Tangential" and centripetal	.....
Bailey	.....	Magmatic, vertical, and derived centrifugal (collapse of "dome")
Bishopp	.....	Magmatic, vertical, and derived centrifugal (toggle-action)

DOES THE VREDEFORT STRUCTURE REPRESENT A METEORITE SCAR?

When, in 1937, Boon and Albritton asked that question, they shared the sense of the lack of collective security felt by geologists in general concerning any published solution of the Vredefort problem—a lack not likely to be supplied by the more recent, contradictory suggestions of Bishopp and Maree—and also recognized the truth of a principle phrased by Shand: "When one is dealing with an extraordinary phenomenon, no possibility is too extraordinary to be worthy of consideration."[30] Boon and Albritton developed the idea that some crustal structures (including a few which had been called "cryptovolcanic") in North America may be the scarring effects of infalls of massive meteorites millions of years ago and then applied the idea to the Vredefort ring-structure. They wrote: "Impact and explosion of a gigantic meteorite could account for the salient structural features of the Vredefort area."[31] Such an infall would explain: (1) the transverse and oblique faults around the margins of the core, (2) the absence of volcanic materials, and (3) the evidence of "unprecedented pressures" and the intense crushing of the rocks of what are here called core and collar.[32] They described their suggestion with extreme brevity. For that reason and also because the suggestion was published in a journal not widely known to the geological and geophysical professions, it has not received the attention it deserves.[33]

Absolute proof that the Vredefort deformation resulted from impact is manifestly impossible. The erosion of probably much more than two hundred million years has utterly destroyed the original relief of the ring-structure and may be expected to have removed all remnants of the assumed meteorite. Hence some direct evidences of infall, such as are given at craters recently formed on the earth, are not to be had.

[30] P. 217 of ftn. 12.

[31] P. 63 of ftn. 2.

[32] P. 64 of ftn. 2.

[33] The present writer is doubtful of one conclusion of Boon and Albritton, namely, that the compressional waves sent horizontally into flat-lying strata from the locus of meteoritic impact should be, as it were, stereotyped in the form of concentric anticlinal and synclinal folds in those beds.

The discussion of a supposed parallel at Vredefort must therefore be theoretical and fearsomely difficult. Reasoning must be partly on the principle of analogy, which cannot be easily guaranteed as true homology, and partly by theoretical deductions, which in the nature of the case cannot be thoroughly verified by experiments. Yet, in the writer's opinion, both analogies and theoretical considerations tend to breed tolerance for the conception that celestial forces, rather than terrestrial forces, were dominant in the development of the ring-structure.

F. G. Watson[34] has listed more than a score of pits made in the earth's crust by infall of (chiefly iron) meteorites, the collisions having been so recent that these craters are still more or less intact. From that list and from data supplied in the corresponding descriptive papers Table 5 has been compiled.

The craters named in Table 5 were wholly or in large part formed by pure percussion, as in the cases represented by rain pits in mud and pits made by speedy balls, bullets, or dud shells in deep clay. That any of these listed craters was much deepened and widened by gaseous explosion is uncertain. L. J. Spencer[35] has found evidence of the volatilization of both silica and nickel-iron at one of the Wabar craters, and special investigation may prove other instances. But the explosive energy of such volatilized materials escapes estimation. Professor H. N. Russell in a letter to D. M. Barringer, Jr.,[36] suggested the likelihood of another kind of gaseous explosion resulting from the fall of large stones and irons at high velocity. Each of these pushes ahead of it a "wad" of greatly compressed air, endowed with explosive power sufficient to

TABLE 5

CRATERS DEFINITELY RECOGNIZED AS DUE TO METEORITIC INFALLS

Locality	Approximate Diameter (M.)	Present Depth below Rim (M.)	Observed Dips (Generally Outward) of Strata in Rim		Evidence of High Temperature Developed by Impact
			Average	Range	
Canyon Diablo, Arizona	1,200	175	30°	0°–90°+	Silica glass
Odessa, Texas	170	5	.....	20 –90 +	.....
Brenham, Kansas	15	3	.....	.....	.....
Campo del Cielo, Argentina (one of many)	53	5	.....	.....	Silica glass
Henbury, Australia (one of many)	220×120	15	.....	.....	Silica glass
Boxhole, Australia	175	16	.....	.....	.....
Dalgaranga, Australia	70	5	.....	.....	.....
Wabar, Arabia (one of several)	100	12	.....	.....	Silica glass
Wabar, Arabia	55	.....	.....	.....	Volatilization of both SiO_2 and iron
Ösel, Estonia (one of six)	100	17	45°	28 –85	.....
Ösel, Estonia	50	3+	.....	35 –50	.....
Ösel, Estonia	55×35	3+	.....	.....	.....
Siberia (one out of ten or more)	50	.....	.....	.....	.....

[34] *Between the Planets* (Philadelphia: Blakiston, 1941), p. 139.

[35] "Meteorite Craters as Topographical Features of the Earth's Surface," *Smithsonian Rept. for 1933*, p. 307 (reprinted from *Geog. Jour.*, Vol. LXXXI [March 1933] pp. 227–42).

[36] *Meteor Crater in Northern Central Arizona* (paper read before the National Academy of Sciences, Nov. 16, 1909; privately printed), p. 22.

drive out much rock debris and thus form a crater.

At each of the three craters named in Table 6 the mass of the (unitary) meteorite and its depth of penetration are known. With that information the velocities of impact can be roughly estimated by making two assumptions: first, that each meteorite was a sphere; second, that Poncelet's formula[37] showing the relation between velocity and depth of penetration for spherical bullets applies in all three cases. Neither of these assumptions can be regarded as correct, but with them a general idea of the respective velocities can be obtained. The results of the calculation are given in column 8 of Table 6. Although the "constants" appearing in the Poncelet formula were determined empirically for bullets fired into the different terrains at the relatively low velocities of cannon fire, it is probable that the computed velocities are higher than those with which the three meteorites struck the ground. That the actual velocities of impact were indeed low is a conclusion readily accepted when allowance is made for the braking effect of the atmosphere on missiles of these small dimensions.

TABLE 6

CALCULATED IMPACT VELOCITIES OF THREE METEORITES

Locality and Reference (1)	Nature of Meteorite (2)	Mass of Meteorite (Kg.) (3)	Radius of Meteorite, Assumed Spherical (M.) (4)	Terrane Struck (5)	Diameter of Crater (M.) (6)	Depth of Penetration (M.) (7)	Computed Velocity at Impact (Km/Sec) (8)
Henbury, Australia (p. 161 of ftn. 34)	Iron	200	0.14	Sandstone, slate, etc.	9	3	2.5+
Paragould, Arkansas (p. 322 of ftn. 35)	Stone	375	0.30	Clayey soil	7	2.5	0.8
Knyahinya, Hungary*	Stone	300	0.27	Clayey soil	1.5	3.5	2.0

* G. P. Merrill, "The Meteor Crater of Canyon Diablo, Arizona; Its History, Origin, and Associated Meteoric Irons," *Smithsonian Misc. Coll.*, Vol. L (1908), p. 491.

D. M. Barringer[38] and colleagues estimated the mass of the Canyon Diablo missile as equal to that of a sphere about 400 feet in diameter. A meteorite of that size and density would strike the ground with a large fraction of the velocity it had when entering the atmosphere. This would be, at minimum, about 10 km/sec. The kinetic energy at impact would therefore be so great that much volatilization of nickel-iron and sandstone would appear inevitable. However, none of the many observers at Canyon Diablo has reported evidence of such volatilization, though the sandstone is clearly shown to have been melted by the heat of collision. In view of all the facts, Barringer[39] concluded that the Canyon Diablo crater was excavated by a swarm of meteoritic irons, great in total mass but so spread out that air resistance reduced the impact velocity well below 10 km/sec. Accordingly, this investigator assigned the excavation of the crater to pure percussion, explosion by volatilized meteorite or rock being practically ruled out.

While some of the kinetic energy was

[37] See C. Cranz, *Lehrbuch der Ballistik* (5th ed.; J. Springer, Berlin: 1925), Vol. I, p. 489.

[38] "Meteor Crater in Northern Central Arizona" (address delivered before the Harvard Club of Boston, Oct. 13, 1926 [typewritten copy]), p. 17.

[39] P. 20 of ftn. 36.

used up in heating, even melting, meteorite and rock, a greater part was employed in excavating, ejection, of debris. Still another fraction was responsible for the remarkably thorough pulverization of a vast quantity of the invaded sandstone. Similar "rock flour" or "rock powder" was developed in the small Henbury and Estonian craters.

At the front of the Canyon Diablo meteorite or swarm of meteorites the pressure must have reached many hundreds of thousands of atmospheres. Under that pressure the particles of the rock powder were, according to a principle established experimentally, shot into the invaded strata horizontally outward and also upward, thus furnishing enough properly directed pressure to deform the strata.[40] In this way the sandstone and limestone were upturned to form a large part of the rim of the Canyon Diablo crater. There the upturning was differential, so that in two places the strata are now vertical. At one point B. C. Tilghman[41] recorded overtilting by about 10° beyond the vertical.

At the largest Estonian crater the beds of the invaded dolomite show maximum upturning at nearly 85°.[42] In a personal letter to the present writer, Dr. M. K. Hubbert describes actual overtilting of the caliche (surface calcareous layer) in the rim of the rim of the Odessa (Texas) meteor crater.

As we now turn attention specifically to the problem of the Vredefort ring-structure, we note at once that, if this does represent a meteorite scar, we have to deal with orders of magnitudes much higher than those that characterized the phenomena at Canyon Diablo. There, as at Vredefort, the terrain struck was a series of nearly horizontal sedimentary rocks. According to the analogy, therefore, the Vredefort strata should have been upturned at average angle much steeper than that found at Canyon Diablo; the crushing and shattering should have been still more intense; volatilization of meteorite and rock should have produced much larger volumes of gas and correspondingly greater effect on the final shaping of the crater. All these effects would be drastic, whether the meteorite were stony or of the nickel-iron class; for simplicity in the following discussion the denser kind of material will be assumed.

On the other hand, the analogies derived from all the recently formed craters are of comparatively little help in accounting for several important features of the Vredefort structure, namely: (1) the remarkable degree of overtilting of the collar strata, (2) the upthrust of the Old Granite core through a vertical distance of something like 15 kilometers, and (3) the intense metamorphism of the Witwatersrand sediments making the broad belt which stretches through the northwest and northeast quadrants of the ring-structure.

Thus the differences between the Arizona and South African structures are both qualitative and quantitative. To understand those differences on the basis of the impact hypothesis, it is necessary to make a number of pure assumptions, the choice of which represents a case of "seasoning to taste," with all that that means in causing mental unease. The procedure is not new. Every one of the other published suggestions as to the origin of the Vredefort ring-structure is

[40] P. 489 of ftn. 37; and also C. Cranz in "Ballistische Kraterbildung," *Gerlands Beitr. zur Geophysik*, Vol. XVII (1927), p. 389.

[41] "Coon Butte, Arizona," *Proc. Acad. Sci. Philadelphia*, Vol. LVII (1905), p. 887.

[42] See W. Kranz, "Krater von Sall auf Ösel, wahrscheinlich 'Meteorcrater,'" *Gerlands Beitr. zur Geophysik*, Vol. LI (1937), p. 50.

similarly based on a hypothesis *ad hoc*. Each rests on unsecured premises essential in the argument. Each has been found wanting. Is it not, then, advisable to give the sixth—impact—hypothesis a hearing, even though absolute proof of its truth is now and in the indefinitely distant future quite improbable? The practical question is: Can the Boon-Albritton idea best explain the phenomena at Vredefort and at the same time avoid the difficulties felt with the five competing hypotheses, all of which regard purely terrestrial forces as responsible for the ring-structure?

Let us continue to be bold and use the Canyon Diablo case as a rough guide in a preliminary test of the impact hypothesis when applied to the African structure. To make the comparison concrete, we shall for the moment, arbitrarily, think of the Arizona meteorite as a nickel-iron sphere with diameter of 100 meters (giving nearly the mass deduced by Barringer and colleagues) and velocity of impact at 10 km/sec (perhaps five times the velocity the "swarm" might have had). The diameter of the Arizona crater, before any slumping enlarged it, was about 1,100 meters; the depth of penetration of the meteorite was here about 400 meters.

At Vredefort the diameter of the assumed crater must have exceeded 50,000 meters; the depth of penetration must have been of the order of 15,000 meters, the bottom of the pad of sediments pushed down ahead of the infallen body reaching 10,000 or more meters still deeper into the earth's crust. The minimum impact velocity may be placed at about 10 km/sec; it may have been two to four or more times greater. Let this meteorite be supposed spherical, composed of nickel-iron, and having initial diameter taken to be, alternatively, 5,000, 10,000, and 20,000 meters.[43]

The ratios of the kinetic energy possessed by the Vredefort sphere to that possessed by the Canyon Diablo sphere would range between 125,000 (5,000-meter sphere striking at 10 km/sec) and 128,000,000 (20,000-meter sphere striking at 40 km/sec). Correspondingly, the pressure at the front of the Vredefort missile would range from nearly 3,000,000 atmospheres to about 55,000,000 atmospheres (as calculated from an equation supplied by Boon and Albritton).[44]

The impact momenta of the assumed 5,000-meter, 10,000-meter, and 20,000-meter spheres at Vredefort would have been, respectively, 125,000, 1,000,000, and 8,000,000 times that of the 100-meter meteorite at Canyon Diablo with the same impact velocity of 10 km/sec. All three ratios would be larger (with maximum at 32,000,000) in proportion with increase of impact velocity in each of the three Vredefort cases. And again we note that at Canyon Diablo the missile was probably a swarm, with much smaller impact velocity and impact momentum than a spherical body of the same tonnage would have.

The ratios of kinetic energies, frontal pressures, and momenta are so large that it does not seem wild to think that the imagined crater at Vredefort had width and depth as great as those deduced from the ring-structure and from the thickness of the sedimentary formations.

[43] Professor Francis Birch has suggested the possibility that this sphere might have been fragmented by impact with air and rock. If so, impact velocity and kinetic energy and impact momentum would all be a little smaller than as stated in the text. Such differences would, however, not be vital in the argument founded on the comparison with the Canyon Diablo case.

[44] "The Impact of Large Meteorites," *Field and Laboratory*, Vol. VI (1938), p. 59.

In one important respect, however, the Canyon Diablo crater could not be homologous with that at Vredefort. Because of the relatively low impact velocity at Canyon Diablo, neither meteorite nor invaded rock was volatilized to a discernible extent; hence there the crater must have been formed by pure percussion. If the impact velocity at Vredefort was anywhere between 10 and 40 km/sec, there must have been volatilization on a big scale and that only a few seconds after the instant of initial impact. Boon and Albritton[45] argue that this volatilization was largely concentrated in the pad of sediments pushed down by the speedy meteorite. In any case it would seem inevitable that the excavation of the Vredefort crater, if of meteoritic origin, was caused by gaseous explosion as well as by the pure percussion of solid meeting solid.

The part played by pure percussion has already been indicated in the case of the Arizona deformation. The gigantic pressure, communicated through the pulverized material of the pad ahead of the Vredefort meteorite, would drive much of this matter into the annulus of rock surrounding the crustal sector directly struck by the huge visitor from space. Therewith the strata of the annulus would be tilted up on all sides, the average angle of tilt, produced by this action alone, being much higher than the average at Canyon Diablo.

With much diffidence the writer will now enter on a brief discussion of that tough problem—the cause or causes for the overtilting of the collar strata. The reader is warned beforehand that no very definite conclusion will be reached; any treatment has to be deductive and deduction is bound to be suspect where sound premises are so few and the phenomenon is outside the range of controlled experimentation.

There is some reason to doubt that gaseous explosion alone would have caused important overtilting. The major explosions at Monte Somma and Krakatoa seem not to have brought about even upturning of the old layers of lava and ash. Neither were the chalk beds uptilted by the powerful explosion of 1916 at the mine crater of La Boiselle, France.[46]

On the other hand, the interval of time separating complete penetration of the imagined meteorite at Vredefort and the gaseous explosion could not have exceeded a few seconds. Is it not possible, therefore, that in this case upturning of the strata was the joint product of both percussion and gaseous explosion? At Canyon Diablo there was strong upturning by pure percussion, apparently unaccompanied by gaseous explosion.[47] At Vredefort the conditions would have been different. Is it possible that there the collar strata were, during and immediately after their violent upturning, overthrown by the enormous "centrifugal" pressure of the expanding gas of explosion?

A second cause of the overtilting at Vredefort is perhaps worthy of consideration. It is suggested by the fact that, after the strong upturning of the collar strata, the Old Granite core was pushed up about 15 kilometers, so that its surface now stands at the same level as the Pretoria beds outside the disturbed annulus. In the attempt to understand

[45] P. 63 of ftn. 44.

[46] P. 307 of ftn. 39.

[47] One of G. K. Gilbert's photographs (see his 1895 Presidential Address to the Geological Society of Washington, p. 20 of pamphlet; reprinted in *Science*, Vol. III [New Ser., 1896], pp. 1–13) seems to show that, when a clay ball was dropped on a flat surface of clay, the layers of this invaded clay were overturned.

this fact more speculation may be ventured.

For reasons given in many books and papers the writer believes it wise to assume that the earth has been cooling since its birth. He also favors the idea that, until recent geological time, our planet has had a true crust, resting on a layer—substratum—of eruptible, vitreous basalt. As a corollary he assumes that, when the Vredefort structure was formed, the earth's crust was thinner than in Tertiary time, hundreds of millions of years later, and had a correspondingly steeper thermal gradient. Being hotter and therefore weaker as well as thinner, the crust at the earlier date was specially prone to yield under isostatic pressure. It seems highly probable that the crust sector struck by the assumed meteorite would have been somewhat weakened by shattering. The impact hypothesis implies that a great load was taken off that sector by the colossal explosion, the area of unloading having a minimum diameter of perhaps 60 kilometers. The question arises as to whether the unloaded sector should have yielded to the unbalanced pressure on the substratum basalt, so that this magma rose, pushing upward the granite core. The question can be rephrased: Assuming such isostatic adjustment, can we best explain both the origin of the deep-seated magmatic body (postulated by some of the competing theories about the Vredefort structure) and the upthrust of the Old Granite overlying that magma? A definitive answer is out of the question, but the line of speculation here suggested seems to ease understanding of the emplacement of thick injections of liquid basalt into the granite core, the injections being regarded as apophysal from the large body of liquid in depth. Perhaps, too, the gravity and magnetic anomalies found by Maree and Simpson have the signs and magnitudes expected if, under the ring-structure, the crust has at the present time an extra thickness of frozen basalt matching the mass of the substratum liquid that had risen into the wounded crust.[48]

If, as supposed, the gaseous explosion left a crater with central depth of something like 15 kilometers and a diameter of about 70 kilometers, the deeper, unwounded part of the sial ("granitic" layer of the earth's crust), endowed with some strength, would be bent upward differentially—most at the central axis of the core and least near the outer limit of the collar. When equilibrium was finally attained, after this regional isostatic adjustment, the collar strata might be expected to be strongly overtilted.

Such overtilting would be additional to any produced by the initial impact and would also be a function of the amount of initial upturning. It seems reasonable to think that, because the penetration by the meteorite was progressive, this upturning was most pronounced in the younger strata of the collar, which, therefore, were carried farthest from the vertical.

A third conceivable cause for the overtilting, based on the volume elasticity of the sediments and Old Granite, can be valued only by the trained elastician. He too will be troubled by uncertainties as to the velocity of penetration by the meteorite and as to the velocities of the

[48] The foregoing explanation of the upthrust of the core granite is, of course, quite unlike that of Nel's (see p. 90 of ftn. 3): that the required pressure was supplied, directly or indirectly, by the younger granite when in the magmatic condition.

It may here be remarked that in the opinion of the present writer the origin of the younger granite and associated nepheline-syenite dikes and veins must still be regarded as a mystery.

shock wave in sediments and Old Granite. He will be dealing with mighty forces of compression and restitution—forces directed both vertically and horizontally. The vertical force of restitution would have been differential, causing maximum upward rebound at the center of the crater. The horizontal force of restitution would also have been differential, with maximum along a level at or near the contact of sediments and Old Granite. Is it possible that the elastician, in spite of the elusiveness of this problem, might give assurance that the play of purely elastic forces set up by impact should have participated to some extent in the overtilting?[49]

One other outstanding feature of the Vredefort structure remains on our list of tests for the impact hypothesis, namely, the intense thermal metamorphism of the Witwatersrand strata in the collar. Figure 3 shows how widespread are its effects on the northwestern half of the ring. We recall, too, Nel's explanation of this differential effect in terms of a supposed wide extension of the younger granite magma, at shallow depth, under both collar and core. Another possibility seems worth consideration. The data of Table 2 show that overtilting of the collar strata is at maximum in the northwest quadrant of the ring-structure. Maree's section (Fig. 2 of the present paper) indicates a minimum of overtilting in the southeast quadrant. Thus, field observations suggest that, if the impact hypothesis is correct, the meteorite did not fall vertically but plunged downward at an angle with the horizontal plane less than 90°, toward the north-northwest. On that side the pressure of impact and also plastic deformation of the collar strata, especially those of the Witwatersrand divisions, would be at maximum. In consequence, these sediments would have been specially heated. Their temperature before the deformation averaged probably as much as 350° C., and it would perhaps be doubled by the heat of impact. Considering the fact that the sediments then carried relatively high percentages of connate water, their thorough recrystallization would be inevitable. That the temperature of the Witwatersrand sediments was high even before the invasion by the younger granite is apparent from Nel's[50] observation that thin apophyses of that granite were not chilled by contact with the sediments.

Special heating of the northwest quadrant of the ring-structure is further suggested by the relatively great abundance of the pseudo-tachylyte in the Old Granite of that quadrant, as indicated in Nel's large map.[51]

CONCLUSIONS

It is clear that the testing of the impact hypothesis when applied to the Vredefort ring-structure is like "making bricks without straw." As in so many geological problems, essential data are hidden in the depths of both time and space. At Vredefort the initial topography, with all its significant details, has been swept away; vital conditions for an objective solution of the structural problem were in depths forever invisible and unreachable. It is a case

[49] Perhaps the inward dips of the collar strata might have been locally increased when thick masses of these rocks, loosened by shattering, slid down with some rotation into the deep crater before this cavity was filled by upthrust of the core granite.

[50] Pp. 69 and 86 of ftn. 3.

[51] Is the asymmetry of the Vredefort collar homologous with that of the vast rim of the lunar Mare Imbrium, the serrate lunar Apennines corresponding to the chaotic pile of upturned rock and rock fragments which may be expected to have been formed at the front of the obliquely falling meteorite at Vredefort?

where deduction from uncertain premises, far outrunning induction from observations in the field, represents the only mode of attack; hence, no one who looks fairly at the meager facts in hand can be expected to have fixed faith in the impact idea. Why, then, labor over it? The answer is multiple. First, the idea should be retained until the resources of the geophysical laboratory have been used for relevant and well-designed experimentation. Second, geologists, the world over, may well follow the lead of Boon and Albritton in the search for more structures which can be reasonably attributed to pre-Recent infalls of giant meteorites; if found, comparison of the respective terranes may conceivably add to the data for Vredefort. Third, the impact hypothesis should be retained because it does seem competent to account best for: (*a*) the spectacular displacement of rock and corresponding colossal energy expended at Vredefort, (*b*) the intensity of the crushing of the collar rocks, (*c*) the shattering of the core granite, (*d*) the upturning of the collar strata, (*e*) the contrast of the structural elements in the northwest and southeast quadrants of the ring-structure, (*f*) the upthrust of the core, (*g*) the injections of gabbroid magma, (*h*) the differential metamorphism of the Witwatersrand sediments, and (*i*) probably the origin of pseudo-tachylyte. In addition, the impact hypothesis has two negative advantages: It does not involve the doming of the Vredefort terrane on an incredible scale, and it does not assume great intrinsic power of upthrust for any deep-seated, isolated body of magma, though the ability of magma to transmit pressure of extraneous origin is held most accountable for the rise of the Old Granite core.

On the other hand, the writer's general argument is affected by some major uncertainties. The chief of these relates to the mechanism suggested for the overtilting of the collar strata. In that mechanism we have imagined the possible co-operation of the initial force of impact, the forces at work during isostatic adjustment, and the elastic forces of restitution. Hardly less disturbing is the question as to whether the collar strata could keep the observed degree of continuity if the short-lived forces of the first and third kind were important in the overtilting. The first and third difficulties call for guidance by the expert elastician. The second calls for relevant laboratory experiments, using models with properly reduced dimensions, such as those recommended by M. K. Hubbert.[52] In this connection it may be noted that E. Öpik, the eminent astronomer, has made appropriate computation and found that in relation to the energy of a massive meteorite, comparable with one of the smaller asteroids, rock matter should behave almost as if it were a true liquid.[53] Evidently, then, the value of the meteorite-scar idea in the Vredefort problem is not to be judged fairly unless due attention is paid to relations and principles quite different from those usually emphasized in theories of crustal deformation.[54]

[52] "Theory of Scale Models as Applied to the Study of Geologic Structures," *Bull. Geol. Soc. Amer.*, Vol. XLVIII (1937), p. 1505.

[53] See "Researches on the Physical Theory of Meteor Phenomena, I. Theory of the Formation of Meteor Craters," *Pub. Observatoire Astron. Université de Tartu (Dorpat)*, Vol. XXVIII (1936), p. 1.

[54] The temporary "craters" in water, milk, and glycerin, made by infalling solid spheres and cinematicographically photographed by A. M. Worthington and R. S. Cole ("Impact with a Liquid Surface, Studied by Aid of Instantaneous Photography," *Phil. Trans. Roy. Soc. London*, Ser. A, Vol. CLXXXIX [1897], p. 137, and Vol. CXCIV [1900], p. 175) are suggestively like the imagined crater at Vredefort.

THE PROBLEM OF THE LIPALIAN INTERVAL[1]

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ABSTRACT

Occasionally sweeping generalizations cling in the literature which are not supported by detailed studies by recent workers. The present paper, based largely upon a study of the literature, is an attempt to draw a coherent picture from the many conflicting statements concerning the pre-Cambrian–Cambrian succession.

It is tentatively concluded that many sediments now classed as Lower Cambrian may have been deposited in pre-Cambrian time, resulting in transition from pre-Cambrian to Cambrian sediments with little or no break in deposition. Evidence is also presented to show that the acquirement of the shell-forming habit may have taken place in a comparatively short time and that this time may be represented by continuous marine deposition.

DEFINITION OF TERMS

The extensively used term "Lipalian" was coined by Charles D. Walcott, who defined it as "the era of unknown marine sedimentation between the adjustment of pelagic life to littoral conditions and the appearance of the Lower Cambrian fauna. It represents the period between the formation of the Algonkian continents and the earliest encroachment of the Lower Cambrian sea."[2]

Walcott accepted W. K. Brooks's hypothesis of the origin of life, and, in order to explain the sudden appearance of the highly evolved Cambrian fauna, he postulated marine deposition only in areas which have never been exposed to observation. The unfossiliferous sedimentary formations which apparently do not long antedate known Cambrian formations were considered by Walcott to be of continental origin. Walcott specifically stated that the long Lipalian interval is "represented by deposition of the great series of pre-Cambrian sedimentary rocks on the North American continent."[3]

An entirely different concept from that expressed by the term "Lipalian interval" is that of the pre-Cambrian (Ep-Algonkian) unconformity. Concerning this unconformity, Walcott wrote, "I have completed my study of the relation of the Cambrian and pre-Cambrian in North America and have concluded that the pre-Cambrian unconformity is universal in all known localities of Cambrian sedimentation and that the depressions were epicontinental, mainly nonmarine, and in no way connected with the subsequent Cambrian sedimentation."[4]

The two concepts, even though in part expressed in the same paper, are not entirely compatible. They should not be considered synonymous. They may not, and in many cases do not, coincide in duration of time or in geologic implication. Walcott has pointed out the presence of sediments of continental origin in Lipalian time, in which case the pre-

[1] A preliminary report of this problem was read before the geology section of the Tennessee Academy of Science, Nashville, November 29, 1946.

[2] "Cambrian Geology and Paleontology," *Smithsonian Misc. Coll.*, Vol. LVII (1914), p. 14. The term "Lipalian" was proposed in a paper presented in 1910 but not published until 1914.

[3] "The Cambrian Faunas of China," *Carnegie Inst. Wash. Pub. No. 54*, Vol. III (1913), p. 29.

[4] P. 31 of ftn. 3.

Cambrian unconformity would be between Lipalian sediments and Cambrian sediments and might represent only a small fraction of Lipalian time.

THE PRE-CAMBRIAN UNCONFORMITY

Geologists have long considered the Lower Cambrian to be separated from the pre-Cambrian by a widespread and profound unconformity. However, restudy of Lower Cambrian sections by recent workers suggests that, in some cases, the unconformity may not be the boundary between pre-Cambrian and Lower Cambrian sediments.

The problem presents a twofold aspect. One phase concerns the classification as Lower Cambrian of excessive thicknesses of sediments conformably underlying formations bearing Lower Cambrian fossils; the other phase concerns the relationship of recognized pre-Cambrian formations to the overlying "Lower" Cambrian formations.

The concept of a universal unconformity at the base of the Lower Cambrian was strengthened by the stratigraphic practice of extending the Lower Cambrian downward from the lowermost fossiliferous formation until a marked unconformity was encountered. A few writers have suggested that these thick unfossiliferous formations may in part be pre-Cambrian. More recently, the National Research Council Committee on Cambrian Stratigraphy has followed the practice of placing the base of the Cambrian at the base of the lowermost fossiliferous formation.[5]

Lower Cambrian sediments and a possible transition from pre-Cambrian to Lower Cambrian rocks can be expected in two areas in North America—the Cordilleran trough and the Appalachian trough.[6]

THE CORDILLERAN TROUGH

In the Cordilleran trough recognized pre-Cambrian igneous, metamorphic, and sedimentary rocks may be directly overlain by Lower, Middle, or Upper Cambrian formations, indicating that the unconformity did not have the same time value everywhere in the trough. In many parts of the area the lowermost fossiliferous formation is conformably underlain by considerable thicknesses of quartzite of which the base may be concealed. In the Nopah–Resting Springs, California section reported by J. C. Hazzard[7] the unfossiliferous formations underlying the zone bearing the Obolella fauna are over 8,000 feet in thickness, more than the entire fossiliferous Lower, Middle, and Upper Cambrian. The committee on Cambrian stratigraphy[8] has since placed the base of the Cambrian 4,000 feet above Hazzard's boundary, thereby classifying approximately 4,000 feet of the basal unfossiliferous sediments as Lower Cambrian.

In the northern part of the trough C. F. Deiss has restudied the classic Cambrian sections of British Columbia and Alberta. The work of Deiss[9] indicates a considerable thickness of unfossiliferous sediments (except for Scolithus, etc.) below the Olenellus zone. In the Nevada area extensive studies by H. E.

[5] "Correlation of the Cambrian Formations of North America," *Bull. Geol. Soc. Amer.*, Vol. LV (1944), pp. 993–1003.

[6] Keweenawan formations of the Lake Superior region, which are pre-Upper Cambrian and which may include sediments deposited in Lower and Middle Cambrian time, are not considered here.

[7] "Paleozoic Section in the Nopah and Resting Springs Mountains, Inyo, California," *Calif. Jour. Mines and Geol.*, Vol. XXXIII, No. 4 (1937), pp. 273–339.

[8] P. 996 of ftn. 5.

[9] "Cambrian Formations of Southwestern Alberta and Southeastern British Columbia," *Bull. Geol. Soc. Amer.*, Vol. L (1939), pp. 951–1026.

Wheeler[10] on the basal Cambrian indicate great thicknesses of quartzite below the lowermost fossiliferous formation. Wheeler classifies this thick unfossiliferous formation as "Lower Cambrian and pre-Cambrian(?)"[11] and very ably criticizes the practice of considering it necessary to extend the base of the Cambrian downward until a marked unconformity is encountered.

The studies by Deiss indicate that a marked angular unconformity exists in the Montana area; but to the northward, in Alberta, much less of the Beltian has been removed by erosion and the two series are generally conformable.[12] In another paper Deiss states, "In the Bow Valley area [Alberta] every known exposure of Cambrian–pre-Cambrian contact exhibits the basal pebbly Cambrian sandstone resting almost conformably on the slightly eroded surface of the Hector shale."[13]

TABLE 1

CORDILLERAN CLASSIFICATIONS

HAZZARD NOPAH RANGE 1937	DEISS PTARMIGAN PEAK 1939	WHEELER PIOCHE DISTRICT 1943	CAMBRIAN COMMITTEE, 1944 NOPAH RANGE
WOOD CANYON FM. 3033'	MT. WHYTE FM. 275'	PIOCHE SHALE 600'	WOOD CANYON FM.
STIRLING QUARTZITE 2,593'	ST. PIRAN SS. 815'	PROSPECT MT. QUARTZITE, 2,000'	STIRLING QUARTZITE
?			
JOHNNIE FM. 2550'	FORT MOUNTAIN 865'	BASE UNKNOWN	
NOONDAY DOLOMITE 1500'	UPPER BELTIAN SHALE		
PRE-CAMBRIAN			

* POSSIBLE UNCONFORMITY
HEAVY LINE INDICATES BASE OF CAMBRIAN

Recent interpretation also suggests that, contrary to the traditionally accepted view, the unconformity separating the pre-Cambrian from formations now classed as Lower Cambrian is in places neither marked nor profound and may even indicate only minor disconformity.

N. E. A. Hinds, who has devoted considerable study to the pre-Cambrian–Cambrian contact of the western United States, states that the Beltian sediments

[10] "Lower and Middle Cambrian Stratigraphy in the Great Basin Area," *Bull. Geol. Soc. Amer.*, Vol. LIV (1943), pp. 1781–1822.

[11] P. 1809 of ftn. 10.

[12] Pp. 1011–12 of ftn. 9.

[13] "Lower and Middle Cambrian Stratigraphy of Southwestern Alberta and Southeastern British Columbia," *Bull. Geol. Soc. Amer.*, Vol. LI (1940), p. 768.

are largely marine and so similar to Early Paleozoic formations that he considers the Algonkian a Paleozoic period.[14] Classifying the Beltian series as Paleozoic would meet with the disapproval of many geologists, including the writer. However, Hinds's studies, together with those of Deiss and others, are significant in indicating the absence in places of a marked break between pre-Cambrian and Cambrian formations.

THE APPALACHIAN TROUGH

Lower Cambrian sediments in parts of the Appalachian trough reveal a situation quite similar to that existing in the Great Basin area of the Cordilleran trough. The classification of the thick unfossiliferous Ocoee and Chilhowee groups has long been a problem.

Arthur Keith[15] mapped (from the base upward) the Snowbird, Hiwassee, and Cochran formations as the Unicoi group, which is the equivalent of James Safford's[16] Ocoee. The Nichols, Nebo, and Murray formations composed the Hampton group according to Keith and with the overlying Hesse or Erwin quartzite were the equivalent of Safford's Chilhowee. Keith placed the base of the Cambrian at the base of the Unicoi. United States Geological Survey practice[17] follows Keith in including the Unicoi in the Cambrian but disagrees with Keith by including both the Cochran and Hesse formations in the Chilhowee group. Charles Schuchert[18] placed the base of the Cambrian at the base of the Hampton group, thereby including all the Unicoi group in the pre-Cambrian.

C. E. Resser considered the problem in several papers and in one paper[19] placed the base of the Cambrian at the base of the Hesse quartzite, classified the Unicoi and Hampton groups as Beltian, and referred the Beltian to the Paleozoic.

The committee on Cambrian stratigraphy[20] placed the base of the Cambrian at the base of the lowermost fossiliferous formation, the Hesse quartzite, and included the underlying formations in the pre-Cambrian.

P. B. King, H. W. Ferguson, L. C. Craig, and John Rodgers,[21] in a detailed stratigraphic study of the formations of northeastern Tennessee, consider the Unicoi, Hampton, and Erwin formations as forming the basal clastic group of Lower Cambrian age. They state that the basal clastic beds range from 4,000 feet to over 7,500 feet in thickness and assign (at Butt Mountain, Tennessee, where the thickest section is given) approximately 1,100 feet to the Erwin formation.[22] They classify the entire group as Lower Cambrian because "the basal clastic group is a conformable sequence which is overlain conformably by the Shady dolomite and rests unconformably on the pre-Cambrian igneous and metamorphic rocks. No fossils except the worm tube, *Scolithus*, have been ob-

[14] "Late Pre-Cambrian Rocks of Western North America," *Pan-Amer. Geol.*, Vol. LXXI (1939), pp. 44–46.

[15] *U.S. Geol. Surv. Folio 143* (1907).

[16] *The Geology of Tennessee* (Nashville: S. C. Mercer, 1869).

[17] "Lexicon of Geologic Names of the United States," *U.S. Geol. Surv. Bull. 896* (1938).

[18] *Stratigraphy of the Eastern and Central United States* (New York: John Wiley & Sons, 1943), pp. 301–8.

[19] "Cambrian System (Restricted) of the Southern Appalachians," *Geol. Soc. Amer. Spec. Paper 15* (1938), p. 2.

[20] Chart I of ftn. 5.

[21] "Geology and Manganese Deposits of Northeastern Tennessee," *Tenn. Div. Geol. Bull. 52* (1944).

[22] Fig. 5 of ftn. 21.

served in the basal clastic group in northeastern Tennessee. Elsewhere the Erwin and its equivalents contain a few fossils of Lower Cambrian age (Butts, 1940, p. 40). As the Hampton and Unicoi are conformable with the Erwin and consist of similar rocks, and as *Scolithus*, like that in the Erwin, occurs as low as the top of the Unicoi, the whole group is probably of Lower Cambrian age."[23]

TABLE 2

APPALACHIAN CLASSIFICATIONS

SAFFORD 1869	KEITH 1907	U.S.G.S. 1937	RESSER 1938	SCHUCHERT 1943	KING, ET AL 1944	CAMBRIAN COMMITTEE 1944
	HESSE	HESSE	ERWIN	HESSE	ERWIN	ERWIN
CHILHOWEE	HAMPTON: MURRAY	CHILHOWEE: MURRAY	BELTIAN: HAMPTON	MURRAY	BASAL CLASTICS: HAMPTON	PRE-CAMBRIAN
	HAMPTON: NEBO	CHILHOWEE: NEBO		NEBO		
	HAMPTON: NICHOLS	CHILHOWEE: NICHOLS		NICHOLS		
	UNICOI: COCHRAN	UNICOI: COCHRAN	BELTIAN: UNICOI	COCHRAN	BASAL CLASTICS: UNICOI	
OCOEE	UNICOI: HIWASSEE	UNICOI: HIWASSEE		HIWASSEE		
	UNICOI: SNOWBIRD	UNICOI: SNOWBIRD		SNOWBIRD		
	IGNEOUS & METAMORPHIC	IGNEOUS & METAMORPHIC				

HEAVY LINE INDICATES BASE OF CAMBRIAN

At many places in the northern Appalachians the nature of the pre-Cambrian–Lower Cambrian contact has not been determined. Frequently, Lower Cambrian formations are marked by an overthrust at the base; in other places the contact is buried. Concerning the situation in western Vermont, C. E. Resser and B. F. Howell state that "most of the strata . . . . are referred to the Lower Cambrian, but in the absence of fossils, definite correlations are not possible. Some of these strata are lithologically similar to, and may represent parts of the pre-Cambrian."[24]

THE LIPALIAN INTERVAL

The sudden appearance of fossils at the beginning of Cambrian time has long been considered one of the most striking aspects of geologic history. So firmly entrenched is the belief in the slowness of evolution that many writers consider the Lipalian period to have extended far back into the pre-Cambrian. The general opinion is well expressed by Charles Schuchert and Carl O. Dunbar who state, "The suddenness of the appearance of armor in several phyla may be more apparent than real, for the Lipalian interval is one of the greatest breaks in the geologic record and probably represents many tens of millions of years dur-

[23] P. 29 of ftn. 22.

[24] "Lower Cambrian Olenellus Zone of the Appalachians," *Bull. Geol. Soc. Amer.*, Vol. XLIX (1938), p. 202.

ing which the evolution of shells was taking place."[25]

The writer believes that evidence presented by the fossils themselves, both in the Lower Cambrian and in later periods, suggests that the period of development of shells was far shorter than is generally believed and was represented by continuous marine deposition.

Two lines of evidence will be considered, namely, the first appearance and the evolution of new forms since the Lower Cambrian and the nature of the Lower Cambrian fauna.

Evolution, when considered in relation to geologic time, has on numerous occasions proceeded with phenomenal swiftness. The first appearance of every new group of fossils is a "sudden" appearance. Unheralded "first appearances" include *Graptozoa* in the Upper Cambrian and *Bryozoa* in the Ordovician. Many other groups could be added to the list. For such groups there is no evidence of a long period during which the shell-forming habit was acquired. On the other hand, there is no doubt that a long period of evolution and development took place prior to their first appearance as fossils. The ability to form a protective covering was acquired during a time when marine deposition was continuously taking place. There is no more necessity for believing that marine deposition was not taking place for some time immediately preceding the appearance of the Lower Cambrian fauna than there is for believing that deposition was not taking place for some time prior to the first appearance of graptolites or bryozoans or other groups.

The ability to form a protective covering may be regarded as an additional characteristic acquired by an already highly evolved organism. Examples are not rare in which an organism has apparently quickly adopted a characteristic which has markedly changed its life habits. The change from a water habitat to a land habitat is such an instance; the change in locomotive habits of reptiles is another. Still further evidence is presented by the rapid evolutionary history of such groups as ammonites in the Jurassic, reptiles in the Triassic, *Graptoloidea* in the Ordovician, and others. The changes become evident only when comparative forms are well preserved as fossils.

Concerning the fauna of the Lower Cambrian, P. E. Raymond[26] points out that the fauna of the Lower Cambrian numbers only about four hundred and fifty-five species, and, although organisms may be highly evolved, the fauna is "simple"[27] compared to that of later periods. The first fauna of the Lower Cambrian reveals an even greater simplicity. Mason[28] states that in the lowermost fauna, the Nevadia zone, only two species are known. Mason's work also shows that the Nevadia fauna, so far, has been described from only one locality—the Palmetto Range (Barrell Spring) section.

When considering the fauna preserved, it is important to consider the nature of the sediments and the chances of preservation of the record. In the Appalachian trough the Nevadia zone is not known although shaly members of formations preceding the Olenellus zone are considered suitable for preservation. Concerning the lower unfossiliferous formations

[25] *Historical Geology* (New York: John Wiley & Sons, 1941), p. 147.

[26] *Prehistoric Life* (Cambridge: Harvard University Press, 1939), p. 23.

[27] P. 28 of ftn. 26.

[28] "Cambrian Faunal Succession in Nevada and California," *Jour. Paleon.*, Vol. XII (1938), pp. 287–94.

of Alabama, Charles Butts writes, "Certainly the character of these deposits, which consist largely of fine water-laid silts, indicates that they were laid down under conditions favorable to Lower Cambrian forms of life and to the preservation of their remains. The universal absence of fossils in these rocks may, therefore, indicate that they were deposited before the advent of the Lower Cambrian fauna, that is, in pre-Cambrian time."[29]

CONCLUSIONS

1. The "Lipalian" interval and the pre-Cambrian unconformity represent two distinct concepts which should not be confused.

2. In some areas pre-Cambrian sediments pass into "Lower" Cambrian sediments with little or no break in deposition.

3. For the sake of uniformity the base of the Cambrian should, for the present, be drawn at the base of the lowermost fossiliferous formation; the practice of extending the Lower Cambrian downward through great thicknesses of unfossiliferous sediments until an unconformity is encountered should be discontinued.

Most geologists consider the coarse, poorly sorted, basal clastics to be of continental origin, while the zones bearing the lowermost fossils are undoubtedly of marine origin. Although there may well have been no actual break in deposition, the transition from continental to marine sediments should be marked by an apparent disconformity where the advancing Lower Cambrian sea re-worked the uppermost continental sediments. The sediments above and below the disconformity should be the same lithologically, but those above should reveal better sorting and the elimination of the finer material. Such a horizon, even though it transgresses time, would provide a definite boundary between sediments deposited by the Lower Cambrian sea and the earlier unfossiliferous formations.

In the Nopah Range, California section, the committee on Cambrian stratigraphy placed the base of the Cambrian at the base of the Stirling quartzite.[30] Hazzard indicated a "possible disconformity" at the base of this formation,[31] although, as noted earlier, he placed the base of the Cambrian some distance below this horizon.

4. Marine deposition may have been taking place for some time preceding the first appearance of fossils, resulting in conditions suitable for their preservation.

5. The period during which animals acquired the ability to form shells may have been relatively short, as indicated by the time required for changes of comparable magnitude in later geologic history.

6. The nature of the first fauna of the Lower Cambrian suggests that organisms did not possess the ability to form shells at the beginning of Cambrian time but quickly acquired the ability when environmental conditions or pressure of competition with other forms of life necessitated rapid change.

7. If statements listed under 4, 5, and 6 above are logical, the concept of the "Lipalian" or "lost" interval is unnecessary.

ACKNOWLEDGMENTS.—Appreciation for many helpful comments and criticisms is expressed to Dr. Louis M. Cline of the University of Wisconsin, although Dr. Cline is in no way responsible for the conclusions presented above.

[29] "Geology of Alabama: The Paleozoic Rocks," *Ala. Geol. Surv. Spec. Rept.* 14 (1926), p. 60.

[30] Chart I of ftn. 5.

[31] See ftn. 7.

BASE OF THE CAMBRIAN SYSTEM[1]

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ABSTRACT

In the western part of the Great Basin region many thousand feet of marine strata lie conformably beneath the earliest Cambrian faunal (time) horizons. Along the trends of transgression these basal deposits disappear progressively until Lower Cambrian and, finally, Middle Cambrian faunas occur in the beds which lie directly upon pre-Cambrian rocks. Thus Cambrian sedimentation began much earlier in some places than in others; and, if we regard a basal unconformity as the lower limit of the Cambrian system, this limit will vary in time from the age of the strata at places where the faunal and physical bases coincide to the earlier but uncertain age of the lowermost stratum conformably underlying the recognized Cambrian. This variable duration of the "Lipalian interval," together with other similar Cambrian relationships, proves the inadequacy of physical criteria as a satisfactory basis for the establishment of time-rock boundaries.

Since faunal correlation generally is much more precise than diastrophism as a time index, the time-stratigraphic units of that portion of the geological record in which diagnostic faunas occur (Paleozoic and higher) should be delimited biologically. In the interest of local and interregional precision of definition and of validity of stratigraphic concept, the base of the biozone of the trilobite genus *Olenellus* is proposed as the base of the Cambrian system.

INTRODUCTION

Recent progress in regional studies on Cambrian stratigraphy has shown conclusively that the conventional treatment of rock units, with regard to their time relationships, is inadequate. H. G. Schenck and S. W. Muller[2] show that lithologic boundaries, including unconformities, play no consistent role in the demarcation of time-stratigraphic units. In the light of the conclusions of these authors, together with regional studies on the early Paleozoic rocks of the southern Great Basin region, it is evident that the lower limit of the Cambrian system can no longer be regarded as a problem to be solved on the basis of local physical relationships.

The upper limit of the Cambrian, like the boundaries between most of the higher units, has been established, in the last analysis, primarily on a faunal basis. Since the Cambrian is the lowermost of the systems generally amenable to more or less precise time definition and since virtually all geologists associate the beginning of Cambrian time with the initiation of correlative faunas (time indices), the need is apparent for a base to this system and a beginning of the Paleozoic era which exhibits a semblance of world-wide time equivalency.

The scope and purpose of this paper does not include a discussion of the original definition of the Cambrian system. This phase of the problem has been adequately covered by M. Grace Wilmarth[3] and by others. It will suffice to point out that the original British type section has long since been abandoned in so far as the Lower Cambrian and its base are concerned.

Nor is it the purpose here to discuss the Lower Cambrian Waucobian series in terms of the strata represented in its type section and their physical relation-

[1] Read before the annual meeting of the Geological Society of America, Chicago, December 27, 1946. Published by permission of the director, Nevada State Bureau of Mines.

[2] "Stratigraphic Terminology," *Bull. Geol. Soc. Amer.*, Vol. LII (1941), pp. 419–26.

[3] "The Geologic Time Classification of the United States Geological Survey Compared with Other Classifications," *U.S. Geol. Surv. Bull. 769* (1925), pp. 92–103.

ships.[4] In view of the structural complexities of the Inyo region, more comprehensive field studies are needed before these rocks can be employed as a satisfactory standard.[5] However, the fact is significant, that, as generally recognized, the Lower Cambrian of Waucoba Springs and elsewhere is characterized by species of the trilobite genus *Olenellus*.

UNCONFORMITIES AS TIME INDICES A FALSE CONCEPT

The classical approach to this problem, as it too often has been with other boundary dilemmas, is the obvious outgrowth of the original catastrophist viewpoint. It is natural that scientists a century and more ago should choose the unconformities to bound their chance type sections. It is further understandable that many of the countless hiatuses throughout the world should be correlated "approximately" with those marking the "boundaries" between the original systems, thus providing a basis for the widespread belief that diastrophism should be the ultimate basis for subdivision of the geological record. It is not the writer's purpose to offer an extended discussion of the role of diastrophism in stratigraphy. Nevertheless, it seems pertinent to invite attention to the altogether too common misconception of the "diastrophism concept" that an unconformity should be generally, if not universally, present (on the continents) at the base of the Cambrian system or that unconformities may be regarded as time-rock division boundaries rather than as merely one of the many expressions, both physical and biological, of more causal phenomena.

The contemporaneity or near-contemporaneity of geographically separated diastrophic events is often falsely assumed. In the absence of fossils many stratal sequences have been assigned to one time unit or another merely because they are conformable with beds belonging to that unit. Such practice, within reason, is often justified, but it is clearly erroneous to assume that hiatuses immediately preceding or following such questionably assigned sequences are sufficiently contemporaneous to be regarded as time indices or division limits. In numerous other cases fossiliferous strata, time equivalents of which are not present in the type sections, are assigned to the underlying or overlying system on the basis of their faunal affinities to one or the other. This practice is logical, and it obviously leads to the ultimate filling of gaps represented by the unconformities at the type sections. As the gaps are so filled, there remains no choice but to select a faunal or time boundary to separate the two systems in question. Otherwise, if physical relationships are employed locally to establish the boundary, strata of a given age may be assigned to the top of one system in some areas and to the base of the superjacent system in others. In consequence, events and phenomena belonging to the early or late portions of geologic periods may be unnecessarily miscorrelated, and thus the very purpose of employing a time scale in geology loses its meaning for an appreciable portion of the record.

Another difficulty in utilizing unconformities as time indices is the fact that the basal strata immediately above a given break may vary in age from place to place. In addition to these arguments,

[4] C. D. Walcott, "Cambrian Sections in the Cordilleran Area," *Smithsonian Misc. Coll.*, Vol. LIII, No. 5 (1908), pp. 185–88.

[5] The pre-Cambrian and early Paleozoic rocks of the Inyo region are currently under study by J. H. Maxon ("Pre-Cambrian Stratigraphy of the Inyo Range" [abstr.], *Proc. Geol. Soc. Amer.*, *1934* [1935], p. 314).

there is the evident fact that no unconformity can constitute a time boundary. Time continued and physical events occurred during the interval represented, although the boundary may be established within the hiatus

RELATIONSHIPS IN THE SOUTHERN GREAT BASIN AREA

To consider the question of the base of the Cambrian system, in the light of the preceding discussion, attention is directed to the regional time-stratigraphic relationships in the southern Great Basin region (Fig. 1).

In the Nopah Range of southeastern California, the excellently described sections of J. C. Hazzard[6] show the zone of *Olenellus* in the upper part of the Wood Canyon formation, as defined by T. B. Nolan,[7] and the lower part of the overlying strata, which are at present without satisfactory designation. Here the genus *Olenellus* occurs through a stratal thickness of approximately 900 feet. Conformably below, an 8,521-foot sequence of unfossiliferous strata separates the *Olenellus* fauna from the pre-Cambrian erosion surface upon which these rocks were deposited. This sequence consists of the Noonday dolomite, the Johnnie formation, the Prospect Mountain quartzite, and the lower 1,900 feet of the Wood Canyon formation, all of which has been referred by Hazzard[8] and Nolan[9] to the Lower Cambrian series because of an absence of an unconformity within. All these pre-*Olenellus* strata, together with several hundred feet of *Olenellus*-bearing beds (over 9,000 feet in all), accumulated in this portion of the Cordilleran geosyncline before the sea advanced southeastward over the margin of the trough to the area now occupied by Frenchman Mountain, near Las Vegas, Nevada.

At Frenchman Mountain the post-Proterozoic sedimentation began only a short time before the regional disappearance of *Olenellus*, for here the trilobites, which appear to be confined to a stratal thickness of 32 feet, are separated from the basal unconformity by only 390 feet (310 feet of Prospect Mountain quartzite and the lower 80 feet of Pioche shale).

Continuing eastward to Western Grand Canyon,[10] *Olenellus* occurs a mere 182 feet above the unconformity; and, according to E. D. McKee,[11] its horizon is finally occupied by the advancing Prospect Mountain facies in the vicinity of Granite Park,[12] about 33 miles eastward. Thus, at an unknown point in the Grand Canyon east of Granite Park, the basal sediments of the Cambrian sea were deposited contemporaneously with the terminus of the *Olenellus* zone in this region; while in the Nopah Range of California, only about 200 miles distant, the highest

[6] "Paleozoic Section in the Nopah and Resting Springs Mountains, Inyo County, California," *Calif. Jour. Mines and Geol., State Mineralogist Rept.*, Vol. XXXIII (1938), pp. 273–339.

[7] "Notes on the Stratigraphy and Structure of the Northwest Portion of Spring Mountain, Nevada," *Amer. Jour. Sci.*, Vol. XVII (5th ser., 1929), pp. 461–72.

[8] Ftn. 6.

[9] "The Basin and Range Province in Utah, Nevada, and California," *U.S. Geol. Surv. Prof. Paper 179-D* (1943), p. 147, and Pl. 40, opp. p. 152.

[10] E. T. Schenk and H. E. Wheeler, "Cambrian Sequence in Western Grand Canyon, Arizona," *Jour. Geol.*, Vol. L (1942), pp. 882–99.

[11] "Cambrian History of the Grand Canyon Region," *Carnegie Inst. Wash. Publ. No. 563* (1945), p. 28.

[12] The questionable suggestion in Fig. 1 that *Olenellus* beds at Granite Park represent a relict facies fauna is offered here merely as an alternative to McKee's (*ibid.*, pp. 135–36) conclusion that ". . . . the sea apparently advanced in steps with pauses between. . . . ." Either explanation serves the purpose of the present study equally well.

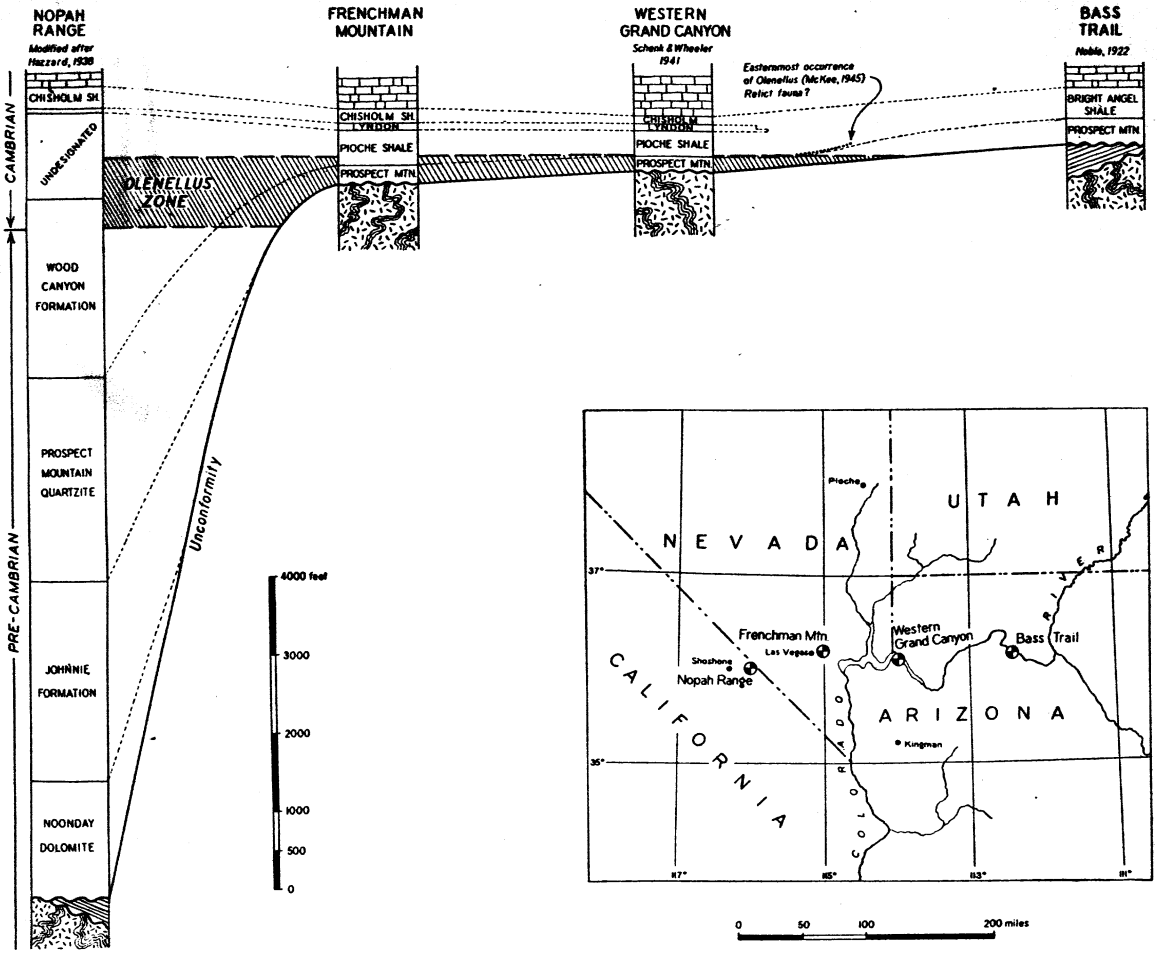

FIG. 1.—Index map and correlation chart showing pre-Cambrian–Cambrian stratigraphic relationships in the southern Great Basin area.

occurrences of this genus are separated from the same unconformity by approximately 9,400 feet!

Under these conditions, can the basal beds above such an unconformity be regarded as a time horizon? Obviously not. And, if not, when and where did the Cambrian sedimentation begin? While the science of geology in general is becoming increasingly quantitative and more precise, *how can we continue to indulge in a concept which regards the base of the Cambrian system as a time horizon but determines that base by physical criteria that are not amenable to interregional correlation?* If we are to continue to regard the beginning of Cambrian time as a "date of reckoning" or a date comparable to the beginning of the Christian era in human history, this system must be given a starting-point in time If we are to regard earliest Cambrian as the time of widespread appearance of faunas amenable to interregional correlation, an appropriate time-stratigraphic (faunal) designation of the base of the system is mandatory. Otherwise, we greatly increase the risk of assigning rocks of a given age to the Cambrian in one region and to the pre-Cambrian in another. To illustrate: in the northern Rocky Mountains an unconformity separates the pre-Cambrian Belt series from Middle Cambrian strata; yet there is no assurance that the uppermost Beltian beds may not be in part contemporaneous with the supra-unconformital Nopah sequence in California. Nor do we have reason to conclude that the lower part of the Sinian system (see p. 158), which conformably underlies the Cambrian of China, differs in age from the higher portion of the Belt series or other strata assigned to the late pre-Cambrian on physical evidence.

PROPOSED BASE OF THE CAMBRIAN SYSTEM

As the most widespread of the earliest faunal (time) horizons admitting of such interregional correlation, the base of the biozone of *Olenellus*[13] is proposed as the base of the Cambrian system. This designation, of course, does not perforce establish the base of the Cambrian at all localities. As in the case of other time-stratigraphic boundaries, identification is contingent upon the local presence of diagnostic faunas and success in determination of the relative ages of the strata in which these faunas make their initial appearances in different regions. For example, in the region discussed herein, the known maximum range of *Olenellus* is found in the Nopah Range, where it extends through approximately 900 feet of strata. In other regions the *Olenellus* biozone is reported to extend through somewhat greater thicknesses. To what extent the differences result from differential rates of sedimentation and to what extent they manifest facies changes or time involved in migration is not known at present. However, until otherwise indicated by additional biostratigraphic data, the lowermost known occurrence of *Olenellus* in this region is tentatively regarded as the base of the biozone of *Olenellus*. Accordingly, the base of the Cambrian system is here tentatively placed approximately 1,900 feet above the base of the Wood Canyon formation in the Nopah Range, as that base was designated by J. C. Hazzard.[14] The pre-

[13] The *Olenellus* biozone includes the *Bonnia* zone, which is also characterized by *Olenellus*.

[14] Pp. 307–12 of ftn. 6, and chart, p. 278. Hazzard admits that the Wood Canyon formation in the Nopah Range includes more beds than were placed in the formation by Nolan at the type locality. The basal 1,535 feet are predominantly arenaceous and perhaps should be included in the underlying

Olenellus-bearing strata of the Wood Canyon, together with the successively underlying Prospect Mountain, Johnnie, and Noonday formations, which aggregate more than 8,500 feet, are assigned to the pre-Cambrian.

The Cambrian Subcommittee of the National Research Council Committee on Stratigraphy[15] points out that L. F. Noble and J. F. Mason believe that the Johnnie formation and the Noonday dolomite are to be regarded as pre-Cambrian. Thus they would place the boundary at the base of the Prospect Mountain ("Sterling") quartzite. This opinion is an appreciable advance in the right direction, but it still offers a solution which is only local in nature and, moreover, not amenable to time correlation. Unless such a lithogenetic boundary is believed to coincide locally with a selected or established faunal boundary, its choice is inconsistent with the facts of time-rock relationships as they are understood at present.

Attention is invited to the fact that the lithogenetic equivalents of the Prospect Mountain quartzite and pre-Cambrian portion of the Wood Canyon formation, as present in the Nopah Range, are not everywhere assignable to the pre-Cambrian. Along the eastward trends of transgression, these shales and quartzites become progressively younger, passing successively from the pre-Cambrian, first, into the Lower Cambrian and, finally, into the Middle Cambrian series. And, conversely, in this easterly direction the pre-Cambrian–Paleozoic boundary courses downward through these lithologic units until it meets the upwardly transgressing regional unconformity.

These general relationships respecting the base of the Cambrian system are by no means confined to the southern Great Basin area. A review of time-stratigraphic relationships in the southern Appalachian region reveals similar conditions. The variable thicknesses of strata below and conformable with those containing the lowermost Cambrian faunal horizons, together with the faunal evidence from younger strata, attest to the geographic variation in the age and duration of the preceding erosion interval; and, as in the Cordilleran region, these differences become even more marked when a comparison is made between the deposits of the Appalachian geosyncline and the bordering areas. R. C. Moore[16] suggests that some of the trough sediments may belong to the late Proterozoic rather than to the Early Cambrian. The Cambrian Subcommittee[17] has removed several of the supra-unconformital formations from the Cambrian, but the base of the Cambrian in each case was placed at a convenient lithologic contact, the contemporaneity of which cannot be proved for any considerable distance.

Further, the same general relationships are found in the development of the Paleocathaysian geosyncline of eastern China.[18] In the border region between Jehol and Hopei provinces, the late pre-Cambrian Sinian system, which has a maximum thickness of more than 17,500

Prospect Mountain ("Sterling") quartzite. In that case the base of the Cambrian would be placed 343 feet above the base of the Wood Canyon formation.

[15] "Correlation of the Cambrian Formations of North America," *Bull. Geol. Soc. Amer.*, Vol. LV (1944), p. 996.

[16] *Historical Geology* (New York: McGraw-Hill Book Co., 1933), p. 126.

[17] Pp. 997–98, Chart 1 and nn. 25, 26, 29 of ftn. 15.

[18] J. S. Lee, *The Geology of China* (London: Thomas Murby & Co., 1939), pp. 67–80 and 211–14.

feet, is conformably overlain by Lower Cambrian strata containing the widespread *Redlichia* fauna. From this maximum the thickness of the Sinian portion of this great continuous sequence diminishes to nothing in parts of southern Shansi and northern Honan. In the southwestern part of the Luliang Range, Cambrian shale rests directly upon the Archean.

Apparently, this problem in eastern Asia was first appreciated in its true light by A. W. Grabau,[19] who separated the Sinian system from the Cambrian on a faunal basis, although he regarded it as the first system of the Paleozoic. Thus, for the first time, a time-stratigraphic base for the Cambrian was established. Through its employment a much truer concept of regional developments in the Paleocathaysian region has been developed than would have been derived if the time-transgressing sub-Sinian unconformity had been employed as the basal unit of the Cambrian in China.

Similarly, in North America the true sequential relationships can be better visualized through the establishment of a time-stratigraphic base for the Cambrian system. Furthermore, the establishment of such a base at the lowermost of the horizons amenable to general interregional correlation would denote a logical separation of the entire geological record into two parts—those strata which do lend themselves to general time-stratigraphic treatment (Cambrian and later) and those which do not (pre-Cambrian).

When the use of a time-stratigraphic base becomes common practice on this continent, the question of general worldwide contemporaneity of the base selected or of the contemporaneity of any time-rock boundary that might prove more appropriate will then be among the detailed problems to be solved in the light of constantly accumulating biostratigraphic data.

[19] "The Sinian System," *Bull. Geol. Soc. China*, Vol. I (1922), pp. 44–88.

ON THE STRUCTURE OF *PSARONIUS SINENSIS* FROM THE OMEISHAN BASALT SERIES IN SOUTHWESTERN CHINA

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The Omeishan basalt is the name given by Y. T. Chao to a thick series of volcanic lavas of the Permian age covering an enormous area in southwestern China. The type locality of the lavas is the Omei Mountain of western Szechuan. Normally this basalt series is underlain by the Lower Permian Yangsin limestone with the leading fossils *Neoschwagerina* and *Wentzelella timorica* and is overlain by the *Gigantopteris* coal series of probably Middle Permian or Saxonian age.[1]

In a paper published in 1942 in the *Bulletin of the Geological Society of China*, the writer described in detail a particularly interesting and well-preserved specimen of Psaronieae discovered in 1941 by C. J. Peng in the Omeishan basalt series in the district of Weining in western Kweichow, probably from the interbasaltic sediments of that series. The species was named *Ps. sinensis* sp. nov. and is probably of Middle Rotliegende age.[2] The present report summarizes the results of the investigation of another specimen (Fig. 1) discovered by the same collector in the same locality. This specimen appears to be a fragment of the so-called "Luftwurzelmantel" of the same species described by the writer in 1942.[3]

[1] H. C. Sze, "Über ein neues Exemplar von *Psaronius* aus dem Omeishan Basalt in Weining (Kweichow) mit besonderer Berücksichtigung des Alters des Basaltes in Südwestchina," *Bull. Geol. Soc. China*, Vol. XXII (1942), pp. 105–31.

[2] *Ibid.*, pp. 123–24.

[3] *Ibid.*, pp. 105–31.

Central portion of the stem.—To go into more detail regarding the anatomy of *Ps. sinensis*, the central vascular system proper of the stem, as has been previously demonstrated and is again shown in this paper, consists of a great number of more or less regularly disposed, elongated steles which have the typical concentric structure, each consisting of a band of wood surrounded by phloem. It follows that a survey of the orientation of the steles may furnish information of great value for the determination of the habit of the plant. Figure 2 has been prepared to demonstrate this plainly. The more minute histology of our specimens has also been worked out with a fair amount of detail. The numerous elongated steles as seen in transverse sections are strongly curved and somewhat concentrically arranged. The xylem band, which is the most conspicuous feature of the stele, is from five to about nine cells in thickness. The tracheids of the metaxylem present nothing out of the ordinary; they are all of the scalariform type so common in the Filicales. The tracheids are nearly isodiametric in transverse section and measure about 70–250 μ in diameter. The xylem parenchyma is not clearly distinguished. Like many other species of *Psaronius*, the protoxylem groups cannot be definitely delimited from the metaxylem. It is often stressed that the siliceous constitution of all the material seems to be unfavorable to the preservation of such delicate structures as the spiral or

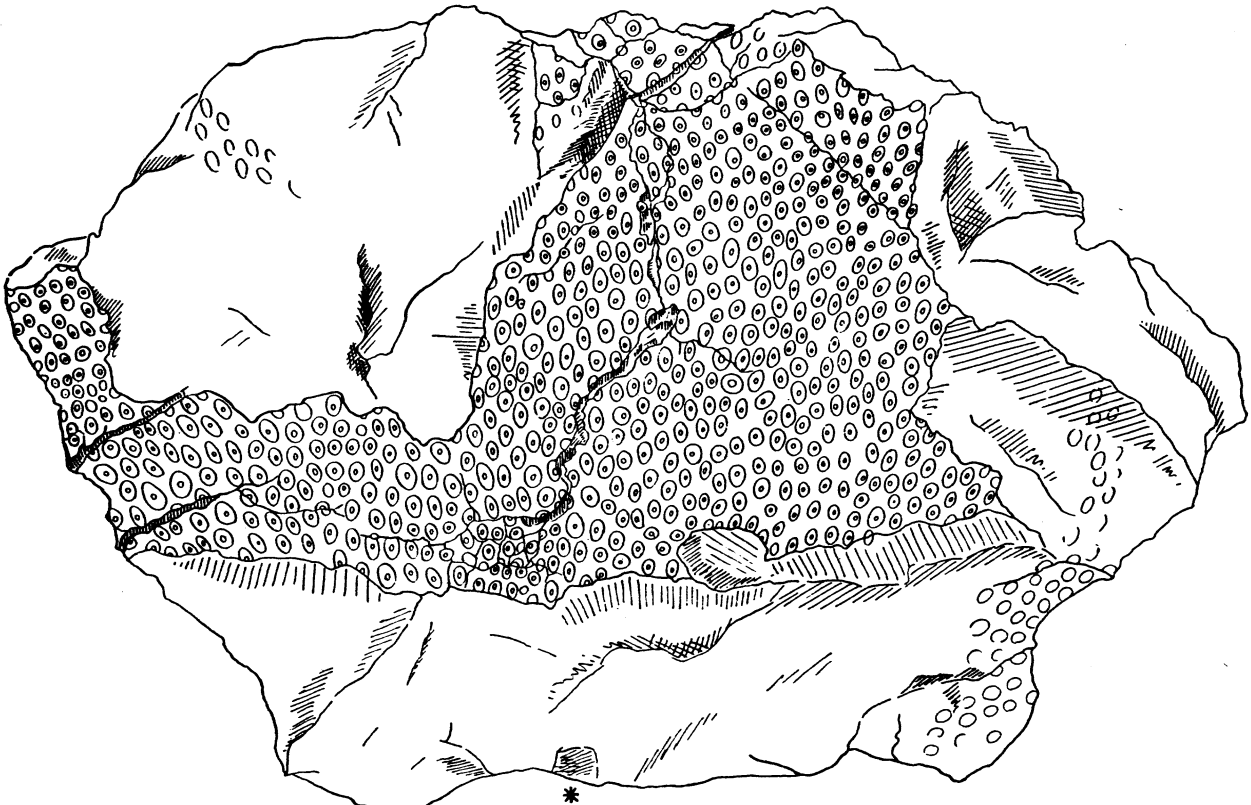


FIG. 1.—Sketch of a transversely cut face of a larger specimen of *Psaronius* in the collection, showing the orientation of the numerous adventitious aerial roots in the root mass. The two specimens of Figures 1 and 2 might belong to the same stem. Note an asterisk on left side of the sketch. This part coincides with that part of Figure 2 marked also with an asterisk. (×7/8.)

annular thickening of the walls of the protoxylem, even though they were originally present. It appears that the protoxylem of our specimen may be endarch and limited to a very small amount, forming a number of small groups on the inner edge of the xylem. On both sides of the xylem the phloem zone is scarcely recognized. It is evident that the phloem tissue was crushed and decomposed during the process of silicification. Also the parenchymatous ground tissue is for the greater part crushed and decomposed, the cavity being filled with silica. In rare instances the parenchyma cells of the ground tissue are still preserved (Pl. I, *4*). The cell walls are thin and delicate. The cells, isodiametrical in transverse as well as in longitudinal sections, measure about 40–90 μ in diameter. There are no irregular bands or patches of sclerenchyma, which are so characteristic of *Ps. infarctus* Ung., to be found in our species. Also the continuous sclerenchyma ring which surrounds the stem and which is very characteristic of *Ps. brasiliensis* Brongn. has not been found in the present stem.

Leaf traces.—Owing to the insufficient and incomplete material available, the position of the leaf traces cannot be definitely determined. On one side of the stem there are two oppositely disposed, strongly curved, horseshoe-shaped bands (Fig. 2, lt_1 lt_2) that may well be interpreted as the outgoing leaf traces. The bands are here more slender and delicate than the remaining steles. The question whether those on the right and left sides of the stem—four similar horseshoe-shaped slender bands (Fig. 2, lt_3–lt_6)—can also be regarded as leaf traces must be left open for the present.[4] Plate II, *3*, illustrates the general aspect of a part of a leaf trace (lt_1 of Fig. 2). The trace consists of a narrow xylem strand which is invested by the rather delicate phloem

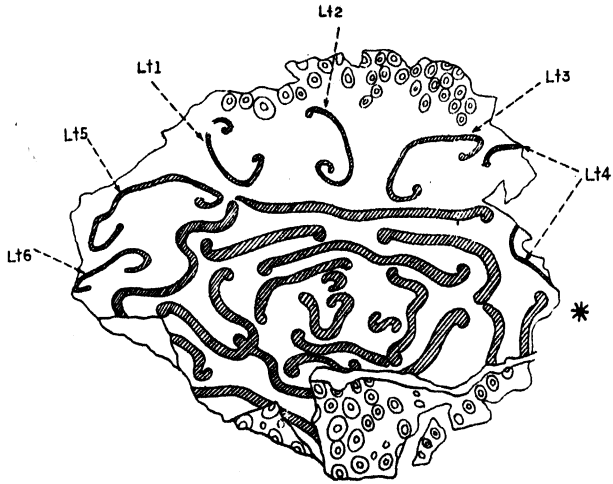


FIG. 2.—Sketch of a transverse section of *Psaronious senensis* Sze showing the orientation of the elongated steles in the central portion of the stem. Note the six smaller and slender, horseshoe-shaped bands (lt_1–lt_6) which might be interpreted as leaf traces. (Natural size.)

[4] *Ibid.*, p. 107.

tissues. The comparatively larger and round cells (Pl. II, *3*) may well be regarded as the characteristic sieve tubes.

Roots.—The mass of the specimen is, as in other species, a mat of adventitious aerial roots, which differs very little within the genus. The vascular structure of the roots is polyarch and is of extreme stellate form. The number of xylem angles is usually five, occasionally also three or four. Phloem and cortex are quite well preserved. The inner cortex is composed of short-celled parenchyma, while the outer cortex is strongly constructed of narrow, thick-walled fibers (sclerenchyma). The thickness of the cell walls of the sclerenchyma measures about 22 μ, while the diameter of the cells is 30–40 μ. As seen in the cross sections, the cells are more or less round, with or without intercellular space. The parenchymatous cells of the inner cortex are sometimes well preserved (Pl. III, *1*). The elements are usually large and thin-walled, measuring about 50–240 μ in diameter. The form of the cells is more or less isodiametrical in cross, as well as in longitudinal, sections. In many instances the parenchymatous cells in the inner cortex were strongly crushed and have perished, forming a wide empty space which has been filled with chalcedony. Between this parenchymatous inner cortex and the central stellate vascular bundles, there is a narrow zone of thick-walled fibers, which might be termed the sclerenchymatous sheath (Pl. II, *1*, *2 sks;* Pl. III, *1*, *2*, *sks*). A few layers of delicate parenchymatous tissues are still to be recognized within this sclerenchymatous sheath. An endodermis has not been clearly distinguished. The central vascular bundles are polyarch, with mostly five or six, occasionally three or four, xylem angles. At the foremost angles of the xylem are situated the smallest elements which no doubt constituted the protoxylem, and longitudinal sections show that they alone are spirally thickened; the remaining elements of the xylem (metaxylem) are scalariform. The tracheids of the metaxylem measure about 40–220 μ in diameter. The phloem is often well preserved. It consists of large, thin-walled, much-elongated elements with oblique end walls. The phloem fills the bays of the stellate xylem, but it does not extend around the ends of the arms, that is, it does not surround the protoxylem. Though sieve plates have not yet been observed, the whole character of the tissue indicates that it is of the nature of phloem.

The adventitious aerial roots that form the closely felted mat around the stem are embedded in and held together by the dense, filamentous tissues that take a radial course (i.e., the tissues run in a horizontal direction). The whole mass is closely matted together to form a practically continuous tissue. Figure 1 illustrates the general aspect of the broad zone of aerial roots. The opinion may here be hazarded, however, that the two fragmentary examples just mentioned (Figs. 1 and 2) might belong to one and the same stem or perhaps even might be split from one and the same block. Although a final decision does not yet seem possible, it would not seem to be too rash to assume the identity of the two examples shown in Figures 1 and 2.

Comparison.—*Psaronius sinensis* is so distinctly different from the several other species of the genus which have been described that there can be no question of its validity. The only species of *Psaronius* previously described that is well enough preserved to permit accurate comparison is *Ps. quadrangulus* Sterzel

from the Middle Rotliegende of Chemnitz in Saxony. According to Hirmer,[5] this species is characterized by

Stamm mit viergliedrig alternierenden Wedelquirlen. In Querschliff zeigen such je vier in einen Kreis gestellte Platten immer alternierend mit de je vier des nächst äusseren und nächst inneren Kreises. Wenn es zur Abgabe des Blattspurstränge (im äussersten Bündelkreis) kommt, werden diese so gebildet dass jede Platte am Rand links und rechts ein schmales Band abtrennt und je zwei Bänder mit den einander zugekehrten Seiten zu einem hufeisenförmigen Blattspurstrang zusammentreten.

According to this description, *Ps. quadrangulus* approaches *Ps. sinensis* in size of the stems and in the orientation of the elongated steles in the central portion of the stem. Although there appear to be no essential differences between the European Middle Rotliegende species and the Chinese type, we are, however, hardly justified in assuming specific identity. Since only a single specimen of the central portion of the stem has so far been discovered, and since, furthermore, the position of the leaf traces has not yet been definitely proved in our species, the writer was inclined during his preliminary studies to identify the Chinese species as new.[6]

From the evidence of constant association, there can be little doubt that the fronds of Psaronieae were of the *Pecopteris* type and identical with those on which fructifications such as *Asterotheca* and *Scolecopteris* were borne. It is of interest to note that a small fragment of a *P. pinna* belonging probably to the groups *P.* (*Asterotheca*) *arborescens* Brongn. and *P.* (*Asterotheca*) *miltoni* Artis sp. was also obtained from our locality.[7] A reconstruction of the probable aspect of *Ps. sinensis* has been given by the writer.[8]

Geological age.—The age of the Omeishan basalt and its overlying *Gigantopteris* coal series has long been a favorite subject for discussion. From the paleobotanist's point of view, so far as the criteria are yet in hand, the basaltic series is evidently bound to the Middle Rotliegende by its petrified fossil *Psaronius*, which has very close relatives in the beds of Middle Rotliegende age of Chemnitz (Saxony), and its overlying *Gigantopteris* coal series is parallel to the Upper Rotliegende horizons in Europe. The writer has discussed this problem at great length in his previous papers.[9] At any rate, there is ample evidence indicating decisively that the flora of the *Gigantopteris* coal series in northern and southern China is Middle Permian or Saxonian in age and cannot be younger, as Dr. T. K. Huang and many Chinese geologists have suggested.[10] Additional evidence in support of this opinion is found in the occurrence of an Upper Permian index fossil *Ullmannia* aff. *bronni* Göpp. in the Talung series in that province. In his previous paper[11] the writer has tabulated the

[5] M. Hirmer, *Handbuch der Paläobotanik*, Band 1 (München, 1927), p. 1.

[6] P. 108 of ftn. 1.

[7] P. 116 and Fig. 5 of ftn. 1.

[8] P. 117 of ftn. 1.

[9] H. C. Sze, "Über das Vorkommen von *Callipteris conferta* Sternb. in der Unteren Shihhotze Serie in Central Shansi," *Bull. Geol. Soc. China*, Vol. XX (1935), p. 572; and pp. 120–25 of ftn. 1.

[10] H. R. Göppert, "Die fossile Flora der permischen Formation," *Palaeontographie*, Vol. XII (1864–65); T. K. Huang, "The Permian Formation of Southern China," *Mem. Geol. Surv. China*, Ser. A, No. 10 (1932); K. Rudolph, "Psaronien und Marattiaceae Denkschr.," *Kaiserl. Akad. Wien*, Vol. LXXVIII (1905); B. Sahni, "Permo-Carboniferous Life Provinces with Special Reference to India," *Current Science*, Vol. IV (1935), pp. 385–90; V. K. Ting and A. W. Grabau, "The Permian of China and Its Bearing on Permian Classification," *Report 16th Internat. Geol. Cong., Washington, 1933* (1934); H. C. Sze, "Über ein Vorkommen von *Ullmannia* aus dem Perm in Kiangsi," *Bull. Geol. Soc. China*, Vol. XX (1940).

[11] P. 124 of ftn. 1.

stratigraphical series of northern and southern China for the later Paleozoic and has attempted thus to show the relations between these formations and also, in a general way, the correlation of the Chinese with the European stages This tabulation is republished as Table 1.

Opinion is by no means agreed upon the possible barrier between the floral provinces of Cathaysia and Gondwana Land at the end of the Paleozoic era. In the paper published in 1942,[12] the writer has stressed the importance of agreeing with Professor Sahni in believing that it is rather the Himalayan geosyncline that must have been an effective barrier between the *Gigantopteris* province and Gondwana Land. Since this paper was published, the writer has received, through the kindness of Professor Sahni, a copy of his important article dealing with the theory of continental drift.[13] It is interesting to see that there is now a consensus among paleobotanists in disagreeing with the view of Ting and Grabau that the Permian basalt highland was a possible barrier between the *Gigantopteris* flora in China and the *Glossopteris* flora in India. Sahni says,[14] for instance:

> It is difficult to say how long a mere basaltic plateau[15] can have remained effective as a barrier between two closely neighbouring floral provinces, connected by land along a front extending all the way southwards as far as Sumatra. . . . . Quite probably the *Gigantopteris* flora will be found to have extended a good deal further west than Ting and Grabau believe.

With his view the writer is in full accord.

TABLE 1

CORRELATION OF CHINESE WITH THE EUROPEAN STAGES

Standard European Stages	Southwest China	North China
Thuringian = Zechstein	Talung Series	? Shihchienfeng Series
Saxonian { Upper Rotliegende / Middle Rotliegende	Lopingian (= Gigantopteris Coal Series) / Omeishan Basalt Series	Upper Shihhotze Series
Artinskian = Lower Rotliegende	Yangsinian { Maokou ls. / Chihsia ls.	Lower Shihhotze Series / Shansi Series (= Upper Yuemenkou Series)
Uralian = Stephanian	Mapingian (= Chuanshanian)	Taiyuan Series (= Lower Yuemenkou Series)
Moscovian = Westphalian	Weiningian	Penchi Series

In addition to the specimens collected from Weining (western Kweichow), a small fragment of *Psaronius* from the Omeishan basalt series of the district Yiliang (eastern Yunnan) has come to the writer through the kindness of Dr. T.

[12] Sze, ftn. 1.

[13] B. Sahni, "Wegener's Theory of Continental Drift in the Light of Palaeobotanical Evidence," *Jour. Indian Bot. Soc.*, Vol. XV (1936), pp. 219–332.

[14] *Ibid.*, p. 326.

[15] In "Wegener's Theory of Continental Drift with Reference to India and Adjacent Countries," *Proceedings of 24th Indian Sci. Cong.* (1937), Sahni points out: "The plateau, however, is not known to have extended further south than the southern border of China. Apart from the improbability of such a plateau remaining effective as a barrier for any appreciable period of geological time even where it did exist, this hypothesis leaves unexplained the anomaly of two climatically distinct floras lying athwart the same latitudes along a front of something like 2,000 miles."

K. Huang of the Geological Survey of China. This fragmentary example is, unfortunately, a mass of polyarch roots without any recognizable stems. In consequence, it is impossible to assign the material to any species, as the roots do not afford specific criteria. The fragment might, however, belong to the species *Ps. sinensis*. The record is of value in that it further outlines the geographic and stratigraphic distribution, not only of the genus *Psaronius*, but also of the entire *Gigantopteris* flora in western China.

The stratigraphical and lithological relationships within the Omeishan basalt series have not yet been fully investigated, but it is certain that there must have been many interbasaltic sediments intercalated in this series. It thus appears clear that the silicified specimens here described were derived from these beds. The same view was expressed by Sahni in a letter to Professor J. S. Lee dated May 29, 1945. In this connection, the writer takes pleasure in citing the words of Sahni, expressed in his general presidential address in 1940, regarding the petrified fossils from the Tertiary Deccan intertrappean beds of India:

> If a lake or river happens to be near by, the ash settled down on the water, forming a sort of volcanic sediment in which the creatures living there find a speedy grave.
>
> But it is an immortal grave. For, through a process that is still largely a mystery to us, the bodies of these plants and animals become imperishably preserved. Particle for particle, cell for cell, the plant tissues are replaced by silica derived from the ash, or from a lava flow that may have overwhelmed the lake; and in the end we are left with an exact replica of the original in hard, indestructible silica.[16]

The same holds good for the present material.

Granting that the greater part of these basaltic lava flows were probably poured under the sea (e.g., in Yunnan, Sikang, and probably also in part of Szechuan),[17] the fact is clear that they should be regarded, at least in eastern Yunnan and western Kweichow, as terrestrial. It thus appears highly probable that, as Sahni has also clearly pointed out in his letter, the basaltic area in these regions would belong to the *Gigantopteris* province itself, instead of forming the barrier between Cathaysia and Gondwana Land. The results of further search in the interbasaltic sediments of the Omeishan volcanic area for many more petrifactions of the *Gigantopteris* flora are therefore eagerly awaited.

[16] B. Sahni, "The Deccan Traps: An Episode of Tertiary Era," *Proceedings of 27th Indian Sci. Cong.* (1940).

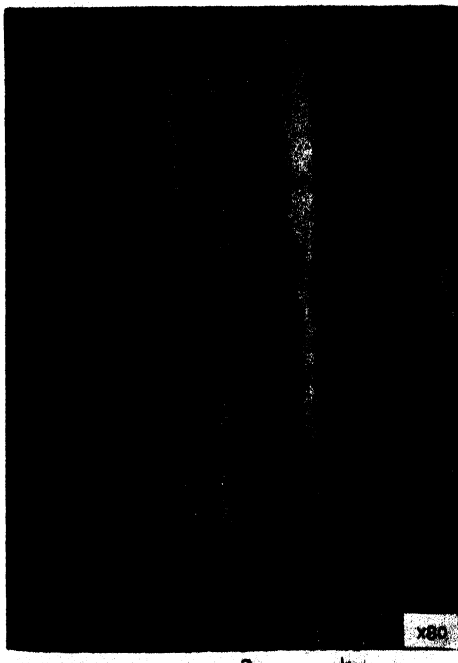
[17] Pp. 124–25 of ftn. 1.

PLATE I

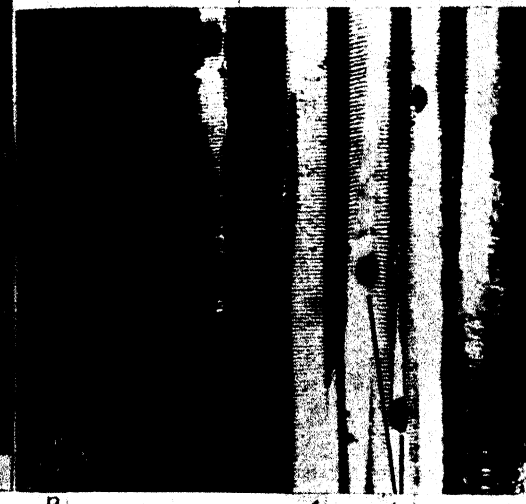
1

5

2



3 k



P 4 k

PLATE II

1
p
sks
ph
x
ph
sks
p

3
X25-6
sv

sks
ph
x
ph
sks
sk
sk
2
X16
pb
sk
r
r
sk
pb

PLATE III

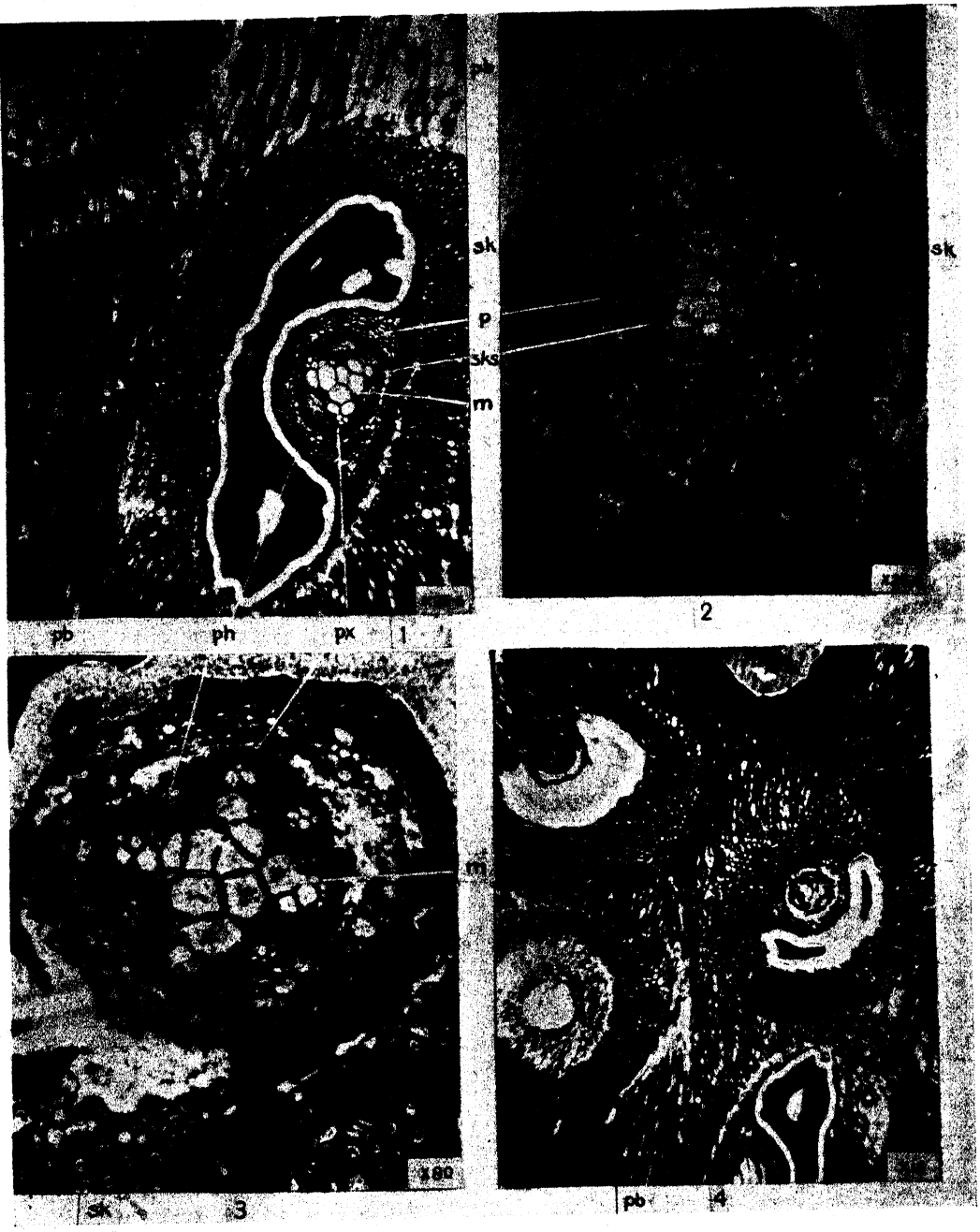


PLATE I

Psaronius sinensis SZE

1, Transverse section of *Psaronius sinensis*, showing the general aspect of the stem and the more or less regularly disposed elongated steles in its central portion. (Natural size.) *2*, Transverse section of a portion of two elongated steles, showing the tracheids of the metaxylem. The position of the protoxylem groups is not well distinguished. The phloem tissues and the parenchymatous ground tissues are fully crushed and decomposed, the cavity being filled with the silica. Xylem-parenchyma is not observed. (×16.) *3*, Longitudinal section of a portion of the stele, showing the scalariform tracheids. Note the occurrence of the irregularly disposed, more or less round, small patches in the tracheids. These small patches (*k*) are brownish in color and may well be interpreted as limonite substances. (×80.) *4*, Longitudinal section of a portion of a stele, showing the scalariform tracheids and the adjacent parenchymatous ground tissues (*p*). Likewise note the small, round, irregularly disposed brown patches in the tracheids (*k*). (×80.) *5*, Transverse section of a portion of the broad zone of the adventitious aerial roots. (Natural size.) The original specimens and slides are in the collection of the Geological Survey of China.

PLATE II

Psaronius sinensis SZE

1, Longitudinal section of an adventitious aerial root. (×80.) *2*, Transverse section of a leaf trace cut through a portion of the central stem. (×25.6.) *x* = xylem (metaxylem); *ph* = phloem; *p* = parenchyma (inner cortex); *sk* = sclerenchyma (outer cortex); *sks* = sclerenchymatous sheath; *pb* = parenchymatous "Bindegewebe"; *r* = the space originally occupied by the parenchymatous tissues of inner cortex, which were for the greater part crushed and decomposed, the cavity being filled with silica; *sv* = presumably sieve tubes of phloem. *3*, Another longitudinal section of a root at a lower magnification. Note the horizontally running parenchymatous "Bindegewebe" (*pb*). (×16.) The original specimens and slides are in the collection of the Geological Survey of China.

PLATE III

Psaronius sinensis SZE

1, Transverse section of an adventitious aerial root, showing the central stellate xylem with three xylem-angles. (×25.6.) *2*, Transverse section of an adventitious aerial root, showing the central stellate xylem with five xylem-angles. (×25.6.) *3*, Transverse section of an adventitious aerial root, showing the central stellate xylem with four xylem-angles. (×80.) *4*, Portion of the section shown in Pl. I, *5*, taken at a higher magnification to show details of the adventitious roots. (×16.) Note the six adventitious roots and the horizontally running parenchymatous tissues ("Parenchymatisches Bindegewebe" [*pb*]). *m* = metaxylem of the xylem; *px* = protoxylem; *p* = parenchyma (inner cortex); *sk* = sclerenchyma (outer cortex); *ph* = phloem; *pb* = parenchymatous "Bindegewebe"; *sks* = sclerenchymatous sheath (a tissue of narrow, thick-walled fibers which surrounds the central stellate vascular bundles and is beneath the parenchymatous inner cortex). The original specimens and slides are in the collection of the Geological Survey of China.

LARGE-SCALE SUPERFICIAL STRUCTURES—A DISCUSSION

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Harvard University

The structural geologist is primarily concerned with the tectonics of the earth. Because of this interest he must be prepared to identify and evaluate superficial structures that simulate those of tectonic origin. Superficial structures have long been recognized not only as annoying obstacles to structural field work but also as valuable criteria of significant processes. The best-known of these disturbances are produced (1) by sliding of beds during deposition; (2) by compaction under load; (3) by expansion due to stripping of overburden; (4) by conversion of anhydrite to gypsum; (5) by gravitational movement under erosion and intensive frost action; and (6) by ice shove.

Nontectonic structures initiated by gravitational movement down the slopes of valleys in beds of gentle dips are described in two recent papers published in England.[1] A comprehensive study of the Northampton Ironstone Field under the direction of the British Geological Survey has yielded, as a by-product, observations on, and analysis of, extensive deformations of relatively large magnitude. Only a sound knowledge of the local geology, intensive field work, and the production of detailed maps and cross sections have enabled the authors to analyze these features and to separate them from those recognized elsewhere in the region as the products of diastrophism. The result is a tightly knit, severely analytical, and convincing study whose importance and general interest to the American geologist prompts this discussion. Dr. S. E. Hollingworth and his colleagues have most kindly granted permission to redraw certain of their illustrations for publication here.

The Northampton Ironstone Field lies north of London in central England. Chalky Boulder-clay, presumably Würm I in age and correlative with the deposits of the Warthe glaciation on the Continent, blankets much of the area. New Drift, considered to be Würm II in age and hence equivalent with the Brandenburg substage of the North German Plains, lies just to the north and west of the area.[2] Beneath the drift, Jurassic rocks are disposed in open folds of gentle dip cut by normal faults, that, like the folds, strike north of west. The Upper Lias clay, 180–210 feet in thickness, is overlain by 150–300 feet of strong sandstone and limestone beds with intercalated weak strata of clay and sand. The economic importance of the Northampton Ironstone, the lower member of this upper group of beds, has warranted the detailed study leading to the recognition of three distinctive types of structure unassociated with the regional structure. These structures, as defined by the authors, are (1) the camber; (2) the gull,

[1] S. E. Hollingworth, J. H. Taylor, and G. A. Kellaway, "Large-Scale Superficial Structures in the Northampton Ironstone Field," *Quart. Jour. Geol. Soc. London*, Vol. C (1944), pp. 1–44; S. E. Hollingworth and J. H. Taylor, "An Outline of the Geology of the Kettering District," *Proc. Geol. Assoc.*, Vol. LVII, Part III (1946), pp. 204–33.

[2] Correlations from H. L. Movius, *The Irish Stone Age* (London: Cambridge University Press, 1942), Fig. 8.

with associated dip-and-fault structure; and (3) the valley bulge.

The term "camber" is introduced by the authors and is applied to an arching of beds over hills, with induced dip toward adjacent valleys. "Cambers result from a lowering of outcropping or near-surface strata. The beds are thus inclined towards the valleys, swathing the hilltops and cloaking or draping the valley sides."[3] This new term is based on the architectural term "camber," defined as "the condition of being slightly arched or convex above."[4] It derives through the Old French *cambre*, "bent, curved," from the Latin *camerare*, "to arch over," and *camera*, "vault" or "arch."

As is well illustrated in Figure 1, a contour map constructed on the contact of the Northampton sand and the Lias clay, the beds involved in the process of cambering include all those above the Lias clay; and, as later brought out, movement also occurred in the clay. The cambers drape the interfluves, and the dip of the beds above the Lias clay conforms to the valley pattern, with an inclination resulting from downhill displacement in excess of 100 feet to the mile. The base of the Northampton beds is seen to fall toward each major valley and even toward the minor tributaries on

[3] Hollingworth *et al.*, p. 3 of ftn. 1.

[4] *Ibid.*, p. 3, n. 1.

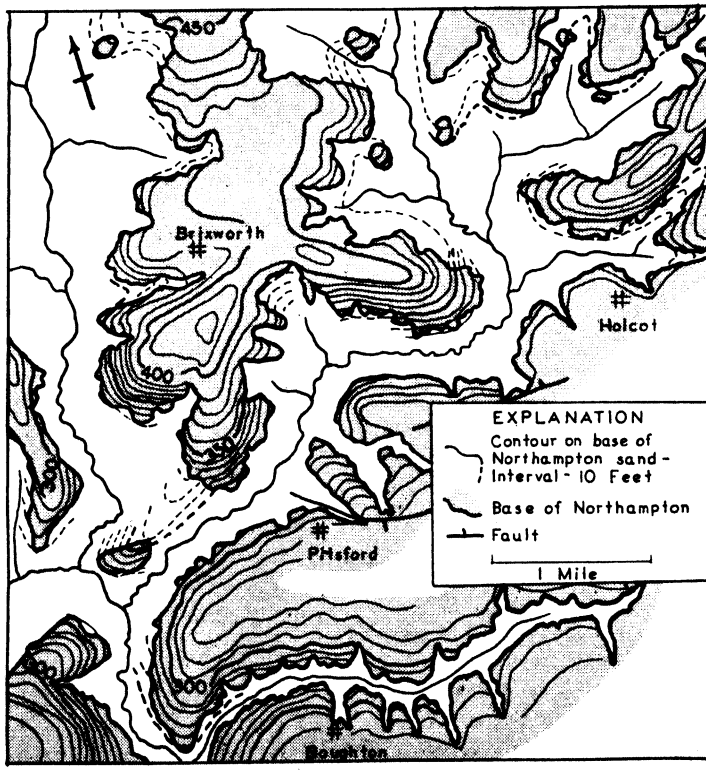

FIG. 1.—Contour map on the base of the Northampton sand to show the relation of cambers to the modern topography. (Adapted from Hollingworth *et al.*, ftn. 1.)

a surface that is obviously but a modified expression of the modern landscape. The authors present detailed sections perpendicular to the surface slopes, which further emphasize this coincidence of "structure" with the present land forms.

Glacial studies prove that, since the deposition of the Chalky Boulder-clay, erosion has removed most of the glacial drift from valleys previously cut into bedrock. Also, records of borings in certain till-choked valleys reveal the presence of cambers deep beneath the surface. The authors thus consider that

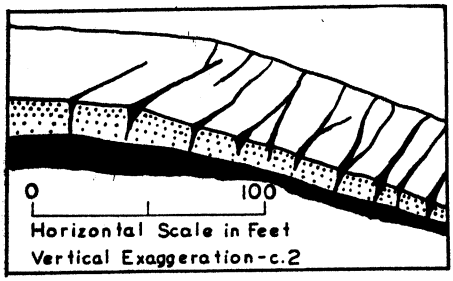

FIG. 2.—Diagram illustrating the relation of gulls to a cambered slope.

most of the cambers antedate the Fourth glacial stage, which was the last ice to cover the area.

The earlier of the two papers under discussion lists three processes involved in the development of cambers: (1) slipping of the competent beds over wet, lubricated clays of the Lias formation; (2) washing out of the clay material below the surface at the spring line; and (3) the outward flow of the Lias clay from beneath the burden of the overlying, more competent beds. The authors are not dogmatic about the dominance of any one of these causes. They leave the impression, however, that additional field study will prove that plastic flow of the Lias clay has been the most decisive of the three processes. In the second paper Hollingworth and Taylor take a stronger stand and consider that cambers result almost completely from the outward and valleyward flow of the Lias clay.

The term "gull" is suggested by the writers for those clefts in the rock which trend normal to the slope of the cambered surface. "Gull" is a local English term for a "gully" or "chasm." Derived from the Old French *gole* or *goule*, it is also used by British quarrymen for widened breaks in rocks with or without infilling.

The gulls are widened fissures or joints in the ironstone and overlying strata, steeply inclined into the hillside, wholly or partly filled with material from above, and normally associated with a camber structure. They tend to parallel each other and the surface contours, although in areas of changes in the direction of slope more complex systems occur. Figure 2, a highly schematic diagram, illustrates the attitude of gulls and their relation to surface slope and cambering.

Gulls are considered to be tensional fractures in the competent beds extended by the downward movement of cambering. The widened gulls result from continued downslope slipping. In calcareous beds the gulls, once initiated by cambering, may be widened by solution induced by water percolating along this new channel.

The authors point out the resemblance of the gulls to ice wedges,[5] but they do not believe that gulling is "explicable in terms of current contraction-crack theories of ice-wedge formation."[6] They suggest, nonetheless, that "further consideration should be given to the part that ground-ice may have played in the devel-

[5] Hollingworth *et al.*, pp. 17 and 18 of ftn. 1.

[6] *Ibid.*, p. 18, n. 1.

opment of the internal structures of a camber."[7] Ice wedging occurs in fine-grained, unconsolidated, but pergelated,[8] sediments of flat-lying areas and produces large-scale polygonal structures.[9] As such, the process is not applicable to the parallel clefts in the consolidated beds of Northampton. Nevertheless, the growth of ice masses as a cause of open clefts in rocks cannot be disregarded. Features similar even in detail to the gulls of Northampton are reported from the vicinity of Prague.[10] Here competent sandstones along valley slopes are fissured parallel to the valley strike. Many of the fractures have been broadened by ice growth within them during the Pleistocene and then infilled from above as the ice melted. Paul Kessler[11] has described open crevices in the Triassic Voltzien sandstone near Saarbrücken. He ascribes these fissures and the implied movement of large sandstone blocks toward the adjacent valleys to the growth of ice masses in incipient openings during a periglacial climate. André Cailleux[12] has reported that the chalk in the region of the lower Seine has been fissured by ground ice in similar periglacial climate. There seems no reason to discard ground ice as a factor in gulling during the periglacial climates that obtained in the Northampton area at various times during the Pleistocene. Tensional fractures, once developed, could have been effectively widened by continued ice accumulation resulting from the freezing of water fed into the gulls by capillary movement from the clays. Adjustment to this pressure would tend to express itself laterally by downslope movement of the intergull blocks. Thus, in addition to being a force in gull formation, ground ice could also have operated in promoting the process of cambering. The infilling of some of the gulls with Chalky Boulder-clay indicates movement of the cambered surfaces in post–Würm I time, even though the cambers may have been initiated at an earlier period. When the southern border of the Würm II ice stood just north and west of the Northampton area on the so-called "York line," the consequent rigorous periglacial climate would have been most effective in producing ground ice. The later cold phases of the last glacial stage would have been less effective.

A common feature in a cambered and gulled structure is a vertical displacement of the beds across the gulls. The amount of movement varies from a few inches to a few feet. The downthrow is almost without exception on the uphill side of the gull, the fault plane dipping into the hill at a high angle. This arrangement is commonly repeated across each of a series of parallel fractures, so that the blocks between consecutive gulls have downhill dips greater than the

[7] *Ibid.*

[8] From the Latin prefix *per*, "throughout," plus the Latin verb *gelare*, "to freeze" (see Kirk Bryan, "Cryopedology—the Study of Frozen Ground and Intensive Frost-Action with Suggestions on Nomenclature," *Amer. Jour. Sci.*, Vol. CCXLIV [1946], pp. 622–42, esp. pp. 635 and 640).

[9] For general reviews of ice wedging see E. de K. Leffingwell, "The Canning River Region, Northern Alaska," *U.S. Geol. Surv. Prof. Paper 109* (1919), esp. pp. 205–42; Stephen Taber, "Perennially Frozen Ground in Alaska: Its Origin and History," *Bull. Geol. Soc. Amer.*, Vol. LIV (1943), pp. 1433–1548, esp. pp. 1510–29; and F. E. Zeuner, *The Pleistocene Period: Its Climate, Chronology, and Faunal Successions* ("Ray Soc.," No. 130 [London: Quaritch, 1945]), pp. 10–13.

[10] Quido Záruba-Pfeffermann, "Periglaziale Erscheinungen in der Umgebung von Prag," *Mitt. d. tschech. Akad. d. Wissensch.*, Vol. LIII (1943), pp. 1–33.

[11] "Über diluviale Frostspalten bei Saarbrücken," *Zeitschr. d. deutsch. geol. Gesellsch.*, Vol. LXXIX (1927), pp. 75–80.

[12] "Fissuration de la craie par le gel," *Bull. Soc. géol. France*, Vol. XIII (5th ser., 1943), pp. 511–20.

general dip of the camber. Such an arrangement of the strata is called by the authors a "dip-and-fault" structure and is diagrammed in Figure 3.

In explaining the dip-and-fault structure the authors concur with C. B. Wedd that it is "the result of movement in a relatively thin, hard and brittle bed intercalated between softer and more plastic strata,"[13] and they add that surface creep may have aided in the forward tilting of the blocks. Further, they state that the "washing out, solution, or

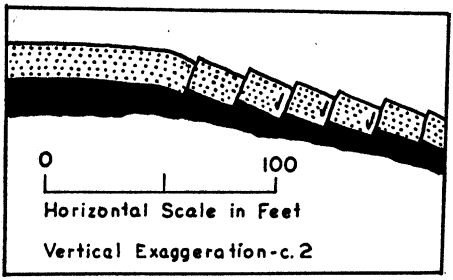

FIG. 3.—Diagram illustrating movement in dip-and-fault structures

flowage of material from beneath a massive jointed bed might be expected to lead to downhill tilting, and this would be accentuated by the consequent displacement of the centre of gravity of the individual block towards its downhill edge."[14] Finally they cite analogous structures resulting from collapse of a competent bed by undermining, the upthrow side being toward the maximum subsidence. Faulting of a similar pattern but on a large scale and of tectonic nature has been described along monoclinal folds by C. B. Hunt,[15] D. L. Babenroth and A. N. Strahler,[16] and H. E. Wright.[17]

The surface movements in competent beds usually produce the Toreva-block landslide type,[18] diagrammatically illustrated in Figure 4. In such a pattern the competent bed assumes a backward dip, and the fault plane dips downslope. The authors state[19] that true landslides, of the Toreva-block type, with accompanying reversals of dip, occur "on actively receding scarps," where cambers are limited in extent. The slopes described by Reiche as affected by Toreva-block movement are relatively steep. Small-scale, frost-induced Toreva-block sliding on high slopes of fissured sandstone near Prague is reported by Zaruba-Pfeffermann.[20] It would appear, then, that the gentle hill slopes prevalent in the Northampton area tend to inhibit normal landsliding and are an inherent part of the environment suitable for the production of dip-and-fault structures.

"Valley bulge" is a term introduced by the authors to describe a feature which in its simplest form is an anticlinal uprising coincident with the lower part of a valley. The structure may, however, occur as a complex fold, a fault block, or a

[13] In G. W. Lamplugh, C. B. Wedd, and J. Pringle, "Bedded Ores of the Lias, Oolites, and Later Formations in England," *Mem. Geol. Surv. Great Britain, Res. xii, Iron Ores* (1920), p. 145 n. (quotation and reference from Hollingworth *et al.*, ftn. 1).

[14] Hollingworth *et al.*, p. 18 of ftn. 1.

[15] "The Mount Taylor Coal Field," *U.S. Geol. Surv. Bull. 860B* (1936), esp. Pl. 19; and "Igneous Geology and Structure of the Mount Taylor Volcanic Field, New Mexico," *U.S. Geol. Surv. Prof. Paper 189B* (1938), esp. Pl. 18.

[16] "Geomorphology and Structure of the East Kaibab Monocline, Arizona and Utah," *Bull. Geol. Soc. Amer.*, Vol. LVI (1945), pp. 107–50, esp. pp. 114, 115, Fig. 2.

[17] "Tertiary and Quaternary Geology of the Lower Rio Puerco Area, New Mexico," *Bull. Geol. Soc. Amer.*, Vol. LVII (1946), pp. 383–456, esp. p. 418, Fig. 10.

[18] Parry Reiche, "Toreva-Block—a Distinctive Landslide Type," *Jour. Geol.*, Vol. XLV (1937), pp. 538–48.

[19] Hollingworth *et al.*, p. 7 of ftn. 1.

[20] See ftn. 10.

narrow horstlike mass bounded by faults. Allied valley-bottom structures exist, in which lateral movement from the sides rather than actual vertical movement appears to dominate.

Figure 5 illustrates in detail the faulting along the axis of a meandering stream valley and the associated up-bulging of the Lias clay. Lateral flow of the Lias valleyward beneath the competent beds is involved in the formation of a bulge and implies a lowering of the strata overlying the zone of greatest outflow. The result in some places is a noticeable downfold or sag along the flank of a bulge. Suggestions of such sags are seen in the sections in Figure 5. The authors' map of known valley bulges, redrawn in Figure 6, shows that bulges are unrelated to regional structure but are coincident with the modern drainage system. The authors present additional detailed maps, both surface and subsurface, which, with accompanying cross sections, drive home this relation between the bulges and the topography and constitute perhaps the most striking phenomena recorded in the two papers.

The incompetent Lias clay has moved upward in the valleys, responding to the load of the strata on the adjacent interfluvial ridges. Similar plastic flow of a weak formation under induced load is not unknown in artificial cuts,[21] and plastic rise of salt to form domes is pointed out by the authors. Although differential specific gravity plays no part in the movement of clay in a bulge as it does in the movement of salt in a dome, the two types of structure are, nonetheless, analogous.

The varied superficial structures of the Northampton area have a common genetic factor in that their development is primarily determined by the behavior, under the influence of erosion, of relatively competent strata that overlie a thick bed of easily deformed clay. The structures appear to have been initiated at different stages in the erosion and development of the present topography. The usual order of development of these features is: bulging in the early stages of the erosion of the valleys, followed by cambering and gulling as erosion cuts down through the Lias clay and isolates the overlying ironstone and accompanying beds on the ridges between valleys.

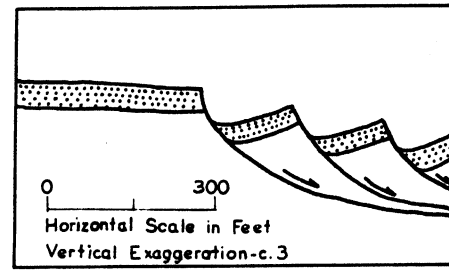

FIG. 4.—Diagram illustrating movement in Toreva-block landsliding.

The authors have ably demonstrated that these structures, in spite of their large scale, are superficial and bear no relation to coexistent forms resulting from regional diastrophism. The forms described—the camber, the gull with associated dip-and-fault structure, and the valley bulge—although in part recognized by the authors' predecessors, particularly the late Beeby Thompson, are for all practical purposes described for the first time.

As far as the writer is aware, similar features have not been described in this country. An unverified exception is the occurrence of apparently minor surface faults in the Yazoo clay of Rankin Coun-

[21] See particularly "Report of the Committee of the National Academy of Sciences on Panama Canal Slides," *Mem. Nat. Acad. Sci.*, Vol. XVIII (1924).

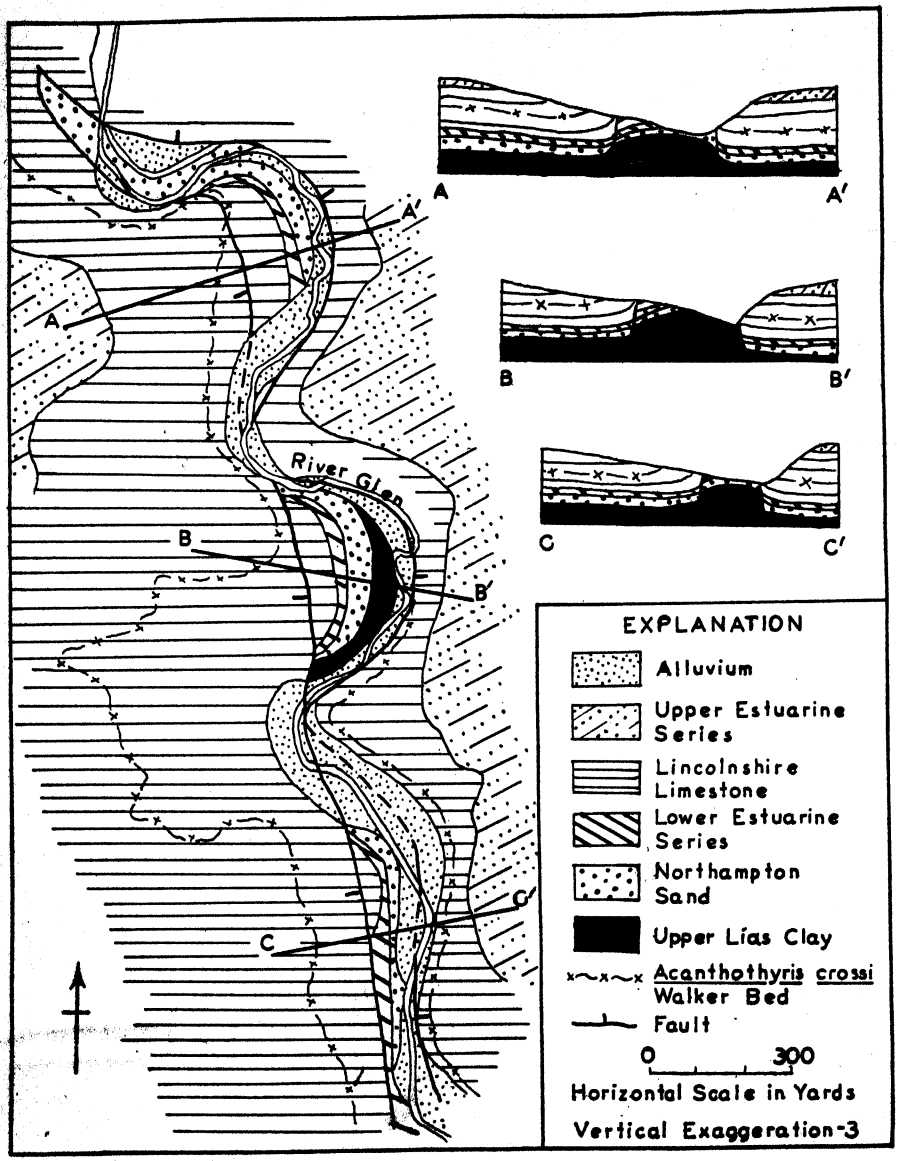

FIG. 5.—Map and sections of Glen Valley between Corby and Creeton to illustrate a valley bulge. (From Hollingworth *et al.*, ftn. 1, after F. B. A. Welch.)

ty, Mississippi.[22] Here a slight upthrow on the downhill side of the fracture, along with the relation of the fissures to local topography and stratigraphy, suggest that they may be comparable to the dip-and-fault structures of Northampton. On the other hand, they may be related to the regional fracture system of the area.[23]

semiaridity of much of the West may be against the presence there of such superficial structures. Yet some resistant strata are found in the southern Coastal Plains; and in the West, where sandstones and plastic shales are common, the climate was not always dry, having been presumably much moister during the Pleistocene stages of glacial advance

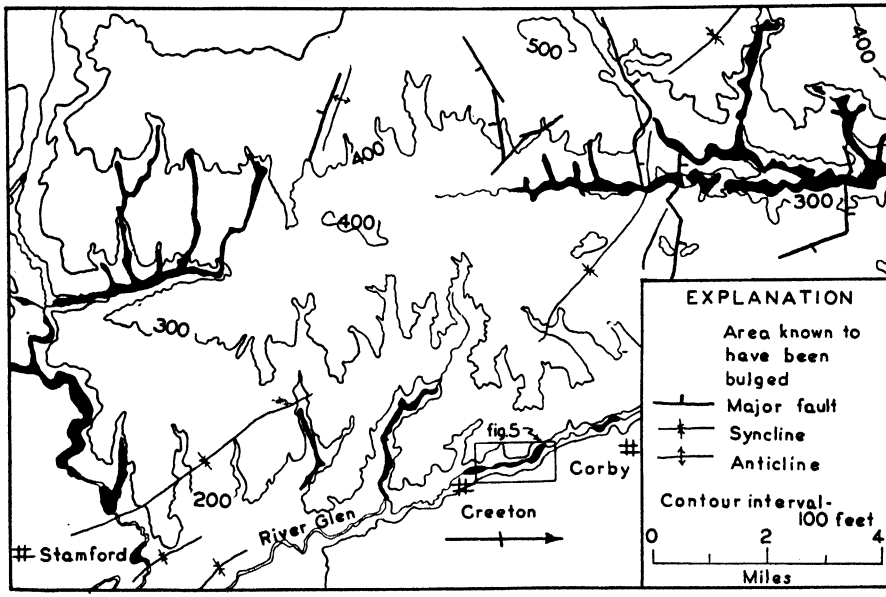

FIG. 6.—Map to show the relation of valley bulges to topography and to regional structure. The area of Fig. 5 is shown by the rectangle in the Glen Valley. (Adapted from Hollingworth *et al.*, ftn. 1.)

It is possible that structures similar to these discussed may be developed to some degree in such regions as the Coastal Plains or in the great areas of Upper Cretaceous beds in western United States. The general lack of competent formations in the Coastal Plains and the

than it is today. In any event, this rarity of reference in the geologic literature to the described forms is not necessarily an index to the actual abundance of these features in the field or their importance to the geologist. Where the requisite stratigraphic sequence and topography obtain, the possibility of their existence should be suspected.

[22] Watson H. Monroe, "Earth Cracks in Mississippi," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XVI (1932), pp. 214 and 215.

[23] H. N. Fisk, "Geological Investigation of the Alluvial Valley of the Lower Mississippi River," *Mississippi River Commission, Vicksburg, Miss.* (1944).

ACKNOWLEDGMENTS.—Professor Kirk Bryan, Harvard University, brought the papers discussed to the writer's attention and encouraged this review of the problem.

REVIEWS

Contributions to Geology, 1944. (University of Texas Publication, No. 4401.) Austin, Tex.: December, 1944. Pp. 567; pls. 55. Paper, $4.25; cloth, $5.00.

This volume is a collection of ten papers (also available separately), eight of which are devoted to various phases of the paleontology of Texas. The other two are concerned with stratigraphy and present geologic background for some of the paleontologic studies.

1. "The Wilberns Upper Cambrian Graptolites from Mason, Texas." By CHARLES E. DECKER. Pp. 13–61; pls. 10; figs. 4. $0.30 separately.

This paper records the occurrence of twenty-four species and varieties of graptolites (eighteen new species, two forms unidentified specifically) referred to five genera, one of which is new; six species of hydroids (four new, one form unidentified specifically) referred to four genera; two genera and species of trilobites an unidentified worm, brachiopod and merostome. These fossils were obtained from a zone considerably older than that occupied by the widespread *Dictyonema flabelliforme* fauna of the Upper Cambrian (Canadian, referred to Lower Ordovician by some) and are stated to include more Cambrian graptolite species and varieties than have been described from all other localities in the world combined. The author includes brief discussions of graptolite morphology, relationships, Cambrian faunas, and correlations and indicates a considerable number of more important references to Cambrian graptolite literature.

2. "Stratigraphy of the Lower Pennsylvanian Coral-bearing Strata of Texas." By F. B. PLUMMER. Pp. 63–76; pls. 3; figs. 1; tables 3. $0.95 with following paper.

The lithology, distribution, thickness, subdivisions, and faunal characteristics of the Marble Falls and Smithwick formations are described. The occurrence of corals at Lower Pennsylvanian collecting localities in north central Texas and in corresponding formations of Oklahoma and Arkansas is shown in tables, and four coral zones are distinguished in this part of the Texas section.

3. "Description of Lower Pennsylvanian Corals from Texas and Adjacent States." By RAYMOND C. MOORE and RUSSELL M. JEFFORDS. Pp. 77–208; pls. 2; figs. 507. $0.95 with preceding paper.

The degree to which Pennsylvanian corals have been neglected is demonstrated by this paper which describes and illustrates sixty-two species (fifty new and five not specifically identified) referred to twenty-two genera (six new and seven not previously recognized in North America), eleven families (three new), and three suborders. A few years ago this entire assemblage would have been classified in about a dozen species and genera on the basis of external characters. Modern studies are based on internal features, however, and require careful cutting of many specimens, but this procedure is neither so exacting nor time-consuming a procedure as is generally supposed.

Corals are abundant at many places in Pennsylvanian rocks, and when they are more accurately and widely known, excellent guide fossils will probably be found among them. This paper is an outstanding contribution initiating a modern study of these fossils, and it will both stimulate, and form a basis for, much-needed additional investigations.

Brief sections on classification and terminology and a very complete bibliography are included.

4. "Smaller Foraminifera in the Marble Falls, Smithwick, and Lower Strawn Strata around the Llano Uplift in Texas." By HELEN JEANNE PLUMMER. Pp. 209–71; pls. 3; figs. 20; tables 1. $0.55 separately.

Assemblages of smaller Pennsylvanian foraminifers are closely related to the types of sediments in which they occur. Specimens from the Marble Falls limestone are mainly calcareous, whereas those from the Smithwick and Strawn shales are mostly arenaceous. This paper describes thirty-eight species and one subspecies (fourteen new and four not identified specifically) referred to twenty-three genera (one new) in eight families. Most of them are too long-ranging stratigraphically to be good guide fossils, but one useful arenaceous species appears to be re-

stricted to the Marble Falls, and several others are not known below the Strawn. Ranges of all species are shown in a table. An excellent section describes the collecting and preparing of material based on Mrs. Plummer's extensive experience in this type of work.

5. "Osteology and Relationships of *Trilophosaurus*." By JOSEPH T. GREGORY. Pp. 273–360; pls. 16; figs. 11. $0.80 separately.

Trilophosaurus buettneri Case was described from a fragment of dentary in 1928 and referred to the Cotylosauria on the basis of the transversely broad teeth which resemble somewhat those of *Diadectes* and *Procolophon*. Extensive quarrying in Howard County, Texas, has yielded a large amount of well-preserved skeletal remains of this genus. The osteology of *Trilophosaurus* is described in great detail, principally on the basis of a nearly complete skeleton. All important osteological details are figured in well-executed text figures and plates. Through this paper *Trilophosaurus* becomes one of the best-known genera of Upper Triassic reptiles.

The skull of *T. buettneri* has a temporal fenestra bounded dorsally by the parietal, posteriorly and laterally by the squamosal, and anteriorly by the postfrontal. Gregory believes that the postorbital extends into the boundary but indicates that the sutures are not sufficiently well shown to be definitive. The fenestra seems to be homologous to the superior temporal fenestra of the Sauropterygia and the Protorosauria and is one of the principal criteria for reference of *Trilophosaurus* to the Protorosauria as proposed in this paper. Other features of the skull and the skeleton confirm this assignment. The new family Trilophosauridae is erected to include the reptile.

The second section of the paper is devoted to an extensive comparison of *Trilophosaurus* with representatives of all major reptilian groups. This is followed by a short discussion of the appearance and habits of the animal. In the final pages a short summary and a complete series of measurements are included. The paper is an important contribution to our knowledge of Triassic reptiles inasmuch as it gives one of the few truly adequate descriptions of a member of the order Protorosauria.

6. "Amphibians from the Dockum Triassic of Howard County, Texas." By H. J. SAWIN. Pp. 361–99; figs. 12; tables 3. $0.25 separately.

This paper is devoted largely to a detailed description of the skull and skeleton of a new species of Dockum stereospondylous amphibian, *Buettneria howardensis* Sawin. The relationships of the species of *Buettneria* and other North American members of the family Metoposauridae are considered in the concluding discussion. Extensive comparisons are contained in comparative tables. The paper adds important details to the knowledge of the genus *Buettneria*, but its most important contribution lies in the badly needed comparative studies.

7. "Decapod Crustaceans from the Cretaceous of Texas." By H. B. STENZEL. Pp. 401–76; pls. 1; figs. 26; tables 2. $0.70 separately.

This paper describes thirty-one species (twenty-one new and one not specifically identified) referred to twelve genera (one new) in nine families. Several of these species are represented by unusually abundant and perfect specimens. There is much new information on relationships, and many details of previously imperfectly known forms are added. An introductory section is devoted to some features of comparative anatomy and stratigraphic ranges of genera, both in Texas and elsewhere, which are tabulated.

8. "An *Amebelodon* Jaw from the Texas Panhandle." By J. T. Gregory. Pp. 477–82; pls. 1; figs. 1. $0.05 separately.

This paper is concerned with the description of an excellently preserved lower jaw of *Amebelodon fricki* and a discussion of the relationships of the genus in which is included a redefinition of the subfamily Amebelodontinae. Dentinal rods have been found by Gregory in a newly described specimen, although they were not observed in the holotype. This, in addition to other evidence, has convinced Gregory that there is no reason for considering the Platybelodontinae and Amebelodontinae, the shovel-tusked mastodonts, to be distinct. He questions seriously the complex classification of Osborn which recognizes three families and eleven subfamilies among the trilophodont type of mastodons. The redefinition of the subfamily Amebelodontinae is clear and defines in concise terms the limits of the group. In discussion of *Platybelodon*, *Torynobelodon* is referred to the genus *Platybelodon*. *Phioma* is regarded as not belonging to the subfamily Amebelodontinae.

9. "Quaternary of the Texas High Plains." By GLEN L. EVANS and GRAYSON E. MEADE. Pp. 485–507; pls. 1; figs. 9. $0.15.

Evans and Meade have made extensive studies of the surface deposits of the Texas

High Plains and have here given an excellent report of their findings and interpretations. The general setting is laid by a brief description of the topography and drainage of the Texas High Plains. Three types of deposits are taken up successively: (1) lake or pond deposits, (2) stream-valley deposits, and (3) wind deposits. The final section of the paper lists the Pleistocene vertebrate localities.

The descriptions and discussion of the lake, or playa, deposits are most detailed and most interesting. Many of the local beds of Blancan and more recent age are shown to be of this type. The origin of the basins of deposition is treated at some length. Subsidence, considered the prime factor by some earlier students, is thought to be of minor importance, and deflation during stages of aridity is indicated as the chief agent producing the basins. The lake deposits appear to have been formed during stages of humid climate. The climate implications of this study are of considerable significance, especially for the relationships of these deposits far from the glaciers and those made at the margins and under the continental ice sheets. Evidence is advanced to show that the Blanco beds are of Nebraskan age. Tule and Spring Creek deposits are considered Middle Pleistocene and the Tahoka clay Wisconsin in age.

Stream and eolian deposits are treated more briefly. It is pointed out that the stream history appears to be closely correlated with climatic cycles and reveals climatic phenomena similar to those shown by the playas. The paper has two important functions: It gives an excellent factual account of the recent geology of the High Plains of Texas and it presents new and interesting interpretations which suggest means of correlating the events of the region with those of other areas of Pleistocene deposition and erosion.

10. "The Blanco Fauna." By GRAYSON E. MEADE. Pp. 509–42; pls. 9; figs. 4. $0.50.

This report is primarily descriptions of an extensive fauna from the Blanco beds in Crosby County, Texas, but it also includes pertinent comments upon the stratigraphic position of the fauna, the nature of its occurrence, and the origin of the Blanco beds. The fauna includes nearly all known members previously described plus seven genera not known from the Blanco heretofore. The locality from which the vertebrates were obtained is described in detail. Meade offers interpretations of the fauna and deposits which contrast markedly with those of various previous students. He believes that the fauna and the nature of the beds, particularly the features indicating climatic conditions during deposition, demonstrate that the Blanco represents a single stage in opposition to Osborn's interpretation of several stages and that the age of the beds is Lower Pleistocene rather than Upper Pliocene. The beds are interpreted as lacustrine rather than as fluviatile, as thought by Matthews.

The descriptions of the members of the fauna are clear, and important specimens are illustrated in the plates. Tables of comparative measurements are included for many of the important groups of vertebrates. This paper adds materially to the knowledge of the Blanco fauna and is challenging in its new interpretations. It should form an excellent basis for comparing the faunas from the Blanco beds with those from other Blancan deposits in North America.

This volume, devoted entirely to the results of strictly scientific research, is effective testimony to the extraordinarily rich and diverse paleontological resources of Texas. It was planned by Dr. E. H. Sellards, former director of the Texas Bureau of Economic Geology, before his retirement. Although it has no direct bearing on economic geology, as that is usually understood, some of these researches will undoubtedly find practical application in the solution of economic problems. Also, these papers are important and permanent additions to essential geological science and are bound to be remembered and consulted long after most contemporary publications on economic subjects have lost their usefulness.

J. MARVIN WELLER
EVERETT C. OLSON

Pennsylvanian Ostracodes of Illinois. By CHALMER L. COOPER. (*Ill. Geol. Surv. Bull. 70.*) Urbana, Ill., 1946. Pp. 177; pls. 21. $1.00.

Cooper's well-illustrated report is the most extensive survey as yet published of the Pennsylvanian ostracodes of this country, covering Morrow to Wabaunsee faunas. Included is a concise stratigraphic summary by J. Marvin Weller of the Pennsylvanian sequence in Illinois, emphasizing the numerous still unsolved problems of correlation. Most of these relate to beds above the Shoal Creek limestone which

would correlate with strata of the Missouri and Virgil series and which occur in scattered outcrops isolated by glacial drift.

In Illinois "ostracodes commonly occur in the upper or marine members of the cyclotherm," although "zones in the underclay limestones and associated beds yield distinctive" faunas commonly considered fresh-water in origin. At least one of the included fresh-water species occurs in the Permian of Kansas. Cooper's discussion of the many zonally restricted species greatly enhances the value of ostracodes as index fossils, even though these short-lived forms are somewhat outnumbered by more hardy species. The early faunas are easily recognized, and few Morrow or Atoka species occur in formations younger than Des Moines. The faunas of the Des Moines series are more like the younger faunas than the older ones. The Missouri series combines species known from either earlier or later strata although a few species are restricted; and the Virgil series combines many short-lived species with older, very hardy, species, many of which continue into the Permian, as well as with a few new species which also continue into Permian strata. The western series and group names have not been adopted in Illinois but were merely used "to conveniently compare Illinois strata with those of the Mid-continent region."

This study shows that most of the Pennsylvanian genera are now well known; of the forty-seven genera found in Illinois only three are new, and few genera are placed in synonomy. Some genera are, however, not yet well established or understood. Cooper believes *Harltonella* Bradfield congeneric with *Waylandella* Coryell and Billings, but the thick downward projecting ridges at the postventral angles of *Harltonella* are very different from the backward projecting spines of *Waylandella*. Instead, *Hartonella* is close to *Coryellites* Kellett, which has not only the same overlap but also the same beaklike ridge on the larger valve. This diagnostic ridge was evidently overlooked by Cooper, for he nowhere mentions it and unfortunately refers to *Coryellites* a number of unornamented species with well-defined dorsal overlap. As for the genus *Carbonita*, Cooper does not follow Scott in referring the Illinois species to the living fresh-water genus *Cypridopsis;* since he is probably correct in retaining the name *Carbonita*, his reasons would have been helpful in a final settlement of this question. Species referred to another living genus, *Candona*, seem to lack the specialized hinge margin of the type species, although possibly they are related to some of the various forms referred to that large and loosely knit genus. The genus *Ectodemites* Cooper seems unacceptable, being founded on a molt of one of the species most closely related to *Amphissites rugosus* Girty, the type of the genus *Amphissites;* furthermore, there seems to be no reason for generically differentiating such forms as *Ectodemites plummeri* Cooper (Pl. XV, Figs. 37–42) and *Amphissites congruens* Cooper (Fig. 43). The synonomy of the often misinterpreted genus *Jonesina*, founded on a European species, has been very unsettled lately, and Johnson's study of type material has been generally overlooked. Recognizing that *J. bolliaformis* is not a true *Jonesina*, Cooper placed it in the genus *Sansabella*. However, near the Wolfcamp-Leonard boundary in northern Texas this species develops the ridges characteristic of *Knoxina* Coryell and Rogatz and should be referred to that genus along with the ancestral Mississippian *Geffienites*.

It is difficult to understand Cooper's denial that the youngiellid type of hingement is present in the genus *Moorites* and in the McAlester *Youngiella* described by Wilson, in view of Wilson's statement that the McAlester species are crenulate and that both he and Knight had observed the minute teeth and sockets extending all along the hinge of *Moorites minutus*, the type species of *Moorites*.

In the past, far too many Pennsylvanian species have been proposed because of failure to recognize immature molts and to make allowance for individual and sexual variation. As a specific character the form-ratio still seems overstressed in this report, especially in the small, plain genera like *Bythocypris*. It is known that in certain living genera, *Loxoconcha*, for example, the male is much more elongate than the female. Cooper's synonymies reduce the number of Pennsylvanian species; but, except for the Bradfield and Payne faunas, we are left in the dark as to whether or not his conclusions are based on the study of type material. Certain conclusions suggest the lack of firsthand knowledge, as in the case of Upson's Funston limestone *Hollinella*, which is removed from the Permian species *H. emaciata* and referred to the Shawnee *H. cushmani*, thus extending the range of the latter well into the Permian. Cooper identifies a Cherokee form with the Shawnee-Wabaunsee *Cavellina fittsi* because of a curved ridge on both which, as he notes, is similar to the ridge

found on the young of many species of *Cavellina*. This ridge, inherited from the ancestral Devonian *Eocytherella*, is much shorter on the Upper Pennsylvanian *C. fittsi* and is found only on very immature specimens not much more than .6-mm. long, as contrasted to the ridged Cherokee forms 1.2-mm. long.

One of the most interesting suggestions made by Cooper, possibly important because bearing on the troublesome question of the orientation of all beyrichias, is that the narrow-frilled hollinellas, being the largest, are females. Usually, it has been assumed that the wide-frilled individuals of this group, forms with the so-called "brood pouch," are the females.

BETTY KELLETT NADEAU

The Journal of Glaciology. Vol. I, No. 1 (January, 1947). Kensington Gore, London, S.W. 7: British Glaciological Society. Pp. 42. 7*s*. 6*d*. per copy.

The British Glaciological Society, founded in 1945, has recently issued the first number of a new scientific magazine—the *Journal of Glaciology*. This new journal is to serve both as the official organ of the Society in the dissemination of glaciologic knowledge and as a general forum for interchange of ideas among students of snow and ice. Its editorial committee consists of Gerald Seligman, president of the Society, Launcelot Fleming, Brian Roberts, and Robert Moss. The new venture seems a timely one, particularly in view of the rapidly growing interest in Arctic and Antarctic regions, which are now becoming far more accessible and important to us than formerly.

This inaugural number contains the proceedings of the Society, several papers presented at its general meetings, various summaries of progress in researches, correspondence, notes, and a bibliography of glaciological literature since 1937. A surprising amount of diverse, vital information is packed within this single issue.

The *Journal of Geology* congratulates the *Journal of Glaciology* on its auspicious start and welcomes it as a promising new member of the geologic fold.

R. T. C.

"Report of the Committee on Marine Ecology as Related to Paleontology," No. 6, 1945–1946. By HARRY S. LADD, Chairman. Division of Geology and Geography, National Research Council, 1946. Pp. 101 (mimeographed). $0.50.

The material presented in this report is of four types: (1) the outline of a proposed treatise on marine ecology and paleontology, (2) announcements of current activities, (3) bibliographic lists, and (4) papers on marine and paleoecology.

The planned treatise is certain to be of great interest to all paleontologists and stratigraphers, and a real opportunity exists to make a major contribution in this field. The best English work on this subject is *Ecological Animal Geography* by Richard Hesse, translated and revised by W. C. Allee and K. P. Schmidt (New York: Wiley, 1937), but this was written for zoölogists, and a somewhat different approach and considerable change in emphasis are required to present the subject matter so that it will have maximum application to paleontology and geology. A work of this type is urgently needed to direct paleontologists' attention to the importance of ecology and to encourage the gathering and recording of pertinent data which are all too commonly overlooked.

As planned, the treatise will consist of two parts, each with an annotated bibliography. The first part will be concerned with marine ecology, presumably largely based on the relations of Recent organisms to their environments. The second part will consist of a series of analyses of selected outstanding regions, one for each period, rather than an attempt to interpret the entire past of all the earth.

The committee desires interested and qualified persons to volunteer their services in the preparation of the various sections. The committee intends to synthesize all contributions into a unified whole so that the volume will not be simply a collection of papers, and individual authorship will not be acknowledged except by the inclusion of a list of contributors in the committee. This plan will work advantageously for the first part, but in my opinion the analyses which will make up the second part should appear under their authors' names, because interpretations of this kind are so subjective that it is important to know who has made them.

The section on current activities is very brief, and only sixteen projects are listed. Observations at Bikini are given by far the greatest emphasis.

About eighty recently published papers are listed—some are briefly annotated. They are partly grouped according to the principal types of organisms. Many of them are concerned with ecology only to a very minor degree.

Three additional bibliographies list (1) Japanese publications that appeared during the war —about forty titles, mostly well abstracted; (2) articles dealing with invertebrate faunas of the Pacific—over one hundred titles, half of them published prior to 1925, mostly briefly annotated; and (3) articles on ecology and paleoecology of New Zealand—about twenty titles dating back to 1913, mostly annotated and some extensively abstracted.

The last section consists of five papers: (1) "Paleoecology in Western Australia," by Curt Teichert; (2) "Methods of Study of Recent Foraminifera," by Fred B. Phleger, Jr.; (3) "Fish Remains in Recent Deposits," by Lore Rose David; (4) "Distribution of Organisms at the Mouth of Wakasa Bay, Japan Sea and Adjoining Continental Shelf," by Hiroshi Niino; and (5) "Life and Environment of the Early Ordovician Sea in Central Texas," by Preston Cloud and Virgil Barnes.

The last paper is by far the most interesting. It is an excellent example of thoughtful and convincing imagination in which past conditions are re-created on the basis of geologic evidence and analogy. Presumably this is the type of interpretation that is desired for the second part of the treatise; and, if other sections can be secured at as satisfyingly high a standard, a most noteworthy result will have been achieved. The reasoning is sound and the conclusions are logical, but an experienced geologist will probably realize that most, if not all, of the observed facts might be interpreted differently. This is not an adverse criticism of the paper but only a warning that our knowledge is so imperfect that few positive conclusions of more than reasonable certainty can be drawn. Much more work of this type is needed, and in time the accumulation of ideas and the testing of theories will lead closer to the results we all desire.

J. MARVIN WELLER

Pennsylvanian Geology of a Part of the Southern Appalachian Coal Field. By HAROLD R. WANLESS. (Geological Society of America Memoir 13, 1946.) Pp. 162; pls. 40.

This volume reports a continuation of Wanless' extensive Pennsylvanian stratigraphic studies that have been described in numerous previous publications, including *Pennsylvanian Correlations in the Eastern Interior and Appalachian Coal Fields* (Geol. Soc. Amer. Special Paper 17). In many ways it is an outgrowth of his northern Appalachian studies. Together with Special Paper 17 it constitutes a more comprehensive description of an important part of the Appalachian Coal Measures than any publication to appear since J. J. Stevenson's "Carboniferous of the Appalachian Basin" (*Bull. Geol. Soc. Amer.*, 1904–6). Although the field work was of reconnaissance nature, Wanless' previous experience; his thorough knowledge of the literature; his painstaking and minute examination of more than four hundred selected stratigraphic sections, some more than 2,000 feet thick, and his application of recent discoveries and modern ideas to the problems of correlation have resulted in a remarkably clear and concise exposition of a very complex succession of strata, reaching a maximum thickness of about 6,000 feet, and their variations throughout an area of more than 7,500 square miles. There can be no doubt that he has brought a great measure of order to a region formerly complicated by at least six different local and unrelated systems of nomenclature and classification. He presents a single group of stratigraphic names and a general classification which is a modification and combination of two Tennessee systems as a framework for his observations but makes no plea for its general adoption.

The major structural features of the southern Appalachian basin are briefly described and shown on a map at the scale of 15 miles to 1 inch; the principal publications dealing with the area are listed; the main stratigraphic subdivisions are described, traced, and correlated; Pennsylvanian deposition is discussed; important unsolved problems are pointed out; and a very complete glossary of stratigraphic names is presented.

Considerably less than one-half of the report consists of ordinary stratigraphic descriptions, and nearly as much space is devoted to the detailed measurements of pertinent portions of more than thirty-five selected stratigraphic sections which illustrate the characteristic features of all but the persistently massive sandstone formations. By far the greater part of Wanless' observations are presented in the form of twenty charts, which show the correlations of a large number of graphic columnar sections, mostly based upon original field measurements. Some of

these correlated sections extend parallel to the structure, and others cut across it at intervals of about 35 miles. The columns are at a scale of about 135 feet to 1 inch and show all principal types of lithology and fossil zones distinguished as bearing marine, brackish, and fresh-water invertebrates and land plants. Each chart is explained and discussed in about a page of fine type. An enormous amount of information is thus compressed into a very modest space.

The Pennsylvanian strata in this region are of Pottsville and early Allegheny age. They are divided into two groups: the Lee, which corresponds to the Pocahontas and New River groups of West Virginia, and a younger unnamed group including beds younger than Kanawha only in the higher mountains of the northern part of the region. Each is separated into several formations ranging in maximum thickness from 100 to 1,800 feet. The Lee formations consist of prominent, massive, conglomeratic sandstones and alternating units of less resistant and more or less shaly strata. Younger formations are heterogeneous units not based on a similar alternation of lithology. Sandstones included in them are generally finer grained and more micaceous than those of the Lee, and they are rarely conglomeratic. Such differences in sandstones and the restriction of a particular type of dark shale to the Lower and Middle Pottsville is widely characteristic and has been noted as far west as Oklahoma.

The cyclical repetition of strata recognized in this region differs from the familiar Illinois type of cyclothem by its greater proportion of coarser clastics which recur in several zones, the occurrence of several coals that may be splits from a single seam, and the rare development of strata with typically marine faunas. Invertebrate fossils, however, were found at a surprisingly large number of places, but most of them suggest brackish or nearly fresh-water conditions. All these differences might be expected in a region of adequate sedimentation close to the source area. Wanless believes that most of the sediment was derived from the old Appalachian land mass, although he recognizes that a smaller part may have been brought in from the Canadian shield.

Stratigraphic correlations substantiate the conclusion, previously based on paleobotanical studies, that the Pottsville section thickens greatly southward from Kentucky. Wanless' correlations, however, demonstrate some formerly unknown details of this thickening which is most notable in the neighborhood of several anticlines and fault zones. Apparently the rate of downwarping in the Pennsylvanian geosynclinal basin differed importantly on either side of these critical zones which became the axes of structural deformation when the region was compressed at the end of the Paleozoic era. His correlations also show the overlap of younger Lee strata onto the Mississippian, westward toward the Nashville dome.

The glossary of about two hundred and fifty stratigraphic names, mostly names of coal seams, is particularly useful because coal-seam terminology was omitted from Wilmarth's *Lexicon*. The type locality (with reference) of each named unit is given, its stratigraphic position is stated, and it is correlated with units known by other names. Also, each is indexed to the text and plates.

The report is illustrated by seventeen plates of excellent photographs and by several formal correlation charts.

J. MARVIN WELLER

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SYMPOSIUM ON PENNSYLVANIAN PROBLEMS

HAROLD R. WANLESS
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Morning and afternoon sessions of Section E (Geology and Geography) of the American Association for the Advancement of Science at St. Louis, Missouri, on March 27, 1946, were devoted to a symposium on the Pennsylvanian. The following papers were presented:

Introduction to Symposium on Pennsylvanian Problems, HAROLD R. WANLESS, University of Illinois, Urbana.

Summary of Classification of the Pennsylvanian Formations of Texas, with Special Reference to the Lower Pennsylvanian of the Llano Region, FREDERICK B. PLUMMER, University of Texas, Austin.

Pennsylvanian Classification and Correlation Problems in North-central Texas, M. G. CHENEY, Anzac Oil Corporation, Coleman, Texas.

Distribution and Correlation of Pennsylvanian Rocks in Late Paleozoic Sedimentary Basins of Northern New Mexico, CHARLES B. READ and GORDON H. WOOD, U.S. Geological Survey, Albuquerque, New Mexico.

Regional Variations in Pennsylvanian Lithology, HAROLD R. WANLESS, University of Illinois, Urbana.

Invertebrates in Pennsylvanian Correlations, J. MARVIN WELLER, University of Chicago, Chicago.

Role of Microfossils in Interregional Pennsylvanian Correlations, CHALMER L. COOPER, Illinois State Geological Survey, Urbana.

Pennsylvanian Floral Zones and Floral Provinces, CHARLES B. READ, U.S. Geological Survey, Urbana.

Plant Microfossils in Correlation of Coal Beds, ROBERT M. KOSANKE, Illinois State Geological Survey, Urbana.

Spore Floras of the Pennsylvanian of West Virginia and Kentucky, AUREAL T. CROSS, University of Cincinnati, Cincinnati, Ohio.

These papers have been assembled for publication, in the hope that they will be thus more useful to geologists working with Pennsylvanian problems.

Linlithgow Library.

INTRODUCTION TO SYMPOSIUM ON PENNSYLVANIAN PROBLEMS

HAROLD R. WANLESS
University of Illinois

Because of their valuable resources of coal, oil, and gas, refractory clays and other ceramic materials, iron ores and other industrial materials; their widespread distribution in the United States; their richly preserved record of the plant and animal life of the period; and their lithologic complexity and variability, Pennsylvanian rocks have been the subjects of many research projects. The "Lexicon of Geologic Names of the United States"[1] records about fourteen hundred names which have been applied to Pennsylvanian units, and this list does not include coal beds, to which at least one thousand additional geographic names have been applied. Two valuable summaries of Pennsylvanian classifications and correlations have recently been published, one representing the co-operative results of a Pennsylvanian subcommittee of the Stratigraphic Committee of the National Research Council, headed by Dr. Raymond C. Moore,[2] and the other a committee of the American Association of Petroleum Geologists, headed by Dr. Monroe Cheney.[3] One might suppose that there would be no necessity for additional programs, such as the present one, on the problems of the Pennsylvanian; but there are still many uncertainties as to correlations in much-studied areas; there are numerous areas, especially in the western states, in which the Pennsylvanian succession is imperfectly known and lumped into thick heterogeneous formations; and many well-represented groups of fossils await detailed study. With the return of peace, many geologists will be able again to turn their attention to such studies as these, and a new generation of college undergraduates and graduate students will be seeking problems suitable for thesis investigations.

The present program has been prepared with the intent of appraising the present state of our understanding of problems and especially to seek out and direct attention to poorly understood areas and to new and promising methods of approach to old problems and groups of fossils, whose detailed study may be expected to contribute to the better understanding of the correlation of, or environments of origin of, certain Pennsylvanian rocks. It was hoped that the chairmen of both committees which have recently worked on Pennsylvanian problems might participate in the program, but Doctor Moore's present brief absence from the country has made it impossible to present a paper he was to give on "Principles of Classification Applied to the Pennsylvanian," an important subject on which complete agreement has not been reached among the principal

[1] M. Grace Wilmarth, *U.S. Geol. Surv. Bull. 898* (1939).

[2] R. C. Moore and others, "Correlation of Pennsylvanian Formations of North America," *Bull. Geol. Soc. Amer.*, Vol. LV (1944), pp. 657–706.

[3] Association Subcommittee on Carboniferous, M. G. Cheney, chairman, "Classification of Mississippian and Pennsylvanian Rocks of North America," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XXIX (1945), pp. 125–69.

students. Dr. Cheney is able to participate in the program, even though he is charged with directing the annual convention of the Association of Petroleum Geologists.

STAGE OF ADVANCEMENT OF REGIONAL STRATIGRAPHIC STUDIES

Little recent study has been made of the Narragansett coal basin, and its relations to other Pennsylvanian areas are poorly understood. It probably resembles in succession and lithology the Maritime Provinces of Canada more than it does other areas in the United States. If any of the metamorphic series of New England which are mapped as Carboniferous are of Pennsylvanian age, their exact age and relations to other unmetamorphosed successions remain to be worked out.

Further detailed studies of the anthracite fields, especially by paleobotanists, are desirable. There are great changes in thickness and facies between the southern and the northern anthracite basins, which, while known, are imperfectly understood. The southern field seems to have a thick Lower and Middle Pottsville section but no Upper Pottsville, while the northern field seems to have a thin Upper Pottsville or Kanawha section but no Middle or Lower Pottsville beds.

Although the type section of the Pennsylvanian is in Pennsylvania, G. H. Ashley[4] and B. Coleman Renick[5] have recently called attention to errors in correlation between Beaver and Allegheny valleys affecting the boundaries of the type Allegheny formation or group and have suggested moving up the basal Allegheny boundary to the base of the Kittanning sandstone.

In Ohio detailed published studies cover only about eight counties—Columbiana and Jefferson in the northern part of the state; Muskingum near the center; and Vinton, Jackson, Scioto, Pike, and Lawrence in southern Ohio. Stratigraphic and faunal studies have been published for the Pottsville[6] and Conemaugh[7] formations. Dr. G. W. White is engaged in studies of the Monongahela of Ohio, and studies of the Allegheny have also been undertaken by M. T. Sturgeon.

In West Virginia and Virginia large county reports cover each county of the Pennsylvanian area, and in West Virginia most beds of a long and very heterogeneous succession have been named. A new contribution to the Pennsylvanian of West Virginia, the succession of spore microfloras by Dr. A. T. Cross, is part of this program. The detailed succession of the early Pennsylvanian underlying the deeper part of the Dunkard Basin is not well described.

The geology of the Maryland Pennsylvanian is well documented in studies published by the Maryland Survey, especially by Dr. C. K. Swartz.

The published literature on the eastern Kentucky coal field is devoted mostly to the tracing of the minable coals, with one paleontologic paper by W. C. Morse.[8] Uniformity in nomenclature between the several river valleys has

[4] "The Pittsburgh-Pottsville Boundary," *Jour. Geol.*, Vol. LIII (1945), pp. 374–89.

[5] "Correlation of the Lower Allegheny–Pottsville Section in Western Pennsylvania," *Jour. Geol.*, Vol. XXXII (1924), pp. 64–80.

[6] Helen Morningstar, "Pottsville Faunas of Ohio," *Ohio Geol. Surv., Bull. 25* (4th ser., 1922).

[7] D. D. Condit, "Conemaugh Formation in Ohio," *Ohio Geol. Surv., Bull. 17* (4th ser., 1912).

[8] "The Pennsylvanian Invertebrate Fauna of Kentucky," *Ky. Geol. Surv.*, Vol. XXXVI (6th ser., 1931), pp. 295–349.

not been attained, and most of the strata between the minable coals have not been named. Data on the unexposed lower Pennsylvanian of the deeper part of the eastern Kentucky field come mostly from inadequate drillers' logs of oil wells. The manner of northwestward and northward thinning and disappearance of the valuable Pocahontas and New River coals is imperfectly known, both in Kentucky and in West Virginia.

A study by the author of the Tennessee coal field, in which its strata are correlated with those in Kentucky, Virginia, and Georgia has recently been published.[9] The faunas of the Pennsylvanian of the southern Appalachian field have been given rather scant attention. Dr. C. H. Summerson[10] has found more than two hundred species of invertebrates in outcrops in West Virginia, Kentucky, and Tennessee. This fauna includes, in addition to the common groups of macrofossils, abundant ostracodes and some foraminifera, conodonts, and holothurians. Fresh-water ostracodes and pelecypods are noteworthy features of the faunas studied by him. A short paper by M. L. Thompson[11] on Ohio material represents all the fusulinid work done in the Appalachian coal field.

The small coal field of northwestern Georgia has been matched quite well with that of adjacent southern Tennessee. Little has been done recently with the Pennsylvanian of Alabama. Different sets of names are applied to coals of the Warrior,[12] Cahaba,[13] and Coosa[14] coal fields, which are in separate fault blocks; and none has been adequately traced across the comparatively barren Plateau coal fields of northeastern Alabama to tie in with the southern Tennessee coal field to the north. Marine faunas are known to exist but are undescribed. Little or nothing is known as to the character of the subsurface Pennsylvanian succession which connects the Alabama fields with those of Arkansas across Mississippi and eastern Arkansas.

In western Kentucky there are too many systems of names initiated in connection with studies of separate quadrangles, and the coals are largely known by numbers not consistently applied to the same coals in different parts of the field and differing from the numbers applied to the same coals in the adjacent Illinois and Indiana fields.

The Indiana field is extensively covered in old volumes by Ashley and his associates, but much new detail about subsurface Pennsylvanian resulting from core drilling for coal or for oil structures has not been assimilated and published. D. W. Franklin[15] has recently published a section for part of southern Indiana, naming several marine limestones which had not previously been differentiated.

Detailed studies of the Pennsylvanian in Illinois are well advanced. The coal division of the Illinois State Geological Survey has for several years collected

[9] "Pennsylvanian Geology of a Part of the Southern Appalachian Coal Field," *Geol. Soc. Amer. Mem.* 13 (1946).

[10] "Some Pennsylvanian Faunas of Tennessee, Eastern Kentucky, and West Virginia" (unpublished Ph.D. thesis, University of Illinois, 1942).

[11] "Pennsylvanian Fusulinids from Ohio," *Jour. Paleon.*, Vol. X (1936), pp. 673–83.

[12] Henry McCalley, "Report on the Warrior Coal Basin," *Ala. Geol. Surv.* (1900).

[13] Joseph Squire, "Report on the Cahaba Coal Field," *Ala. Geol. Surv.* (1890).

[14] A. M. Gibson, "Report upon the Coosa Coal Field," *Ala. Geol. Surv.* (1895).

[15] "Pennsylvanian Stratigraphy of Part of Southern Indiana," *Trans. Ill. Acad. Sci.*, Vol. XXXVI (1944), pp. 85–92.

samples at 1–5-foot intervals from drilling oil tests taken by members of its staff and correlated with drilling-time logs and electric logs. The results have been a vastly increased knowledge of the Middle and Lower Pennsylvanian, where these are buried in the deeper part of the basin.[16] There remains uncertainty as to the relations of the youngest Pennsylvanian in some parts of the Illinois Basin, the position corresponding to the Missouri-Virgil boundary, and the correlation of the isolated high Pennsylvanian rocks of Webster and Union counties in western Kentucky, south of the Rough Creek fault, with the Illinois section.

In Michigan, where the Pennsylvanian covers a large area but is deeply buried by glacial drift, detailed studies by W. A. Kelly[17] of the one small area of good outcrops have been published. The age, lithology, and succession of most of the Michigan Basin Pennsylvanian are, however, poorly known.

Recent detailed studies by Lewis M. Cline and others in Iowa have resulted in tracing numerous thin marker beds of the Upper Des Moines and Lower Missouri groups south across Missouri to eastern Kansas[18] and eastward to tie in with the Illinois section.[19] The higher Missourian beds[20] remain to be matched with beds in Illinois, and strata of Lampasas and Morrow age, if any, have not yet been worked out.

In Missouri the outcropping Pennsylvanian strata have been carefully traced by H. Hinds and F. C. Greene[21] and by others to tie in with both Iowa and Kansas successions. The thickened Lower Pennsylvanian subsurface section of the Forest City Basin has not been so accurately subdivided. Although well described from several diamond-drill records,[22] the Lampasas–Des Moines boundary remains somewhat uncertain.

G. E. Condra[23] has subdivided the outcropping Upper Pennsylvanian of Nebraska about as minutely as has been done anywhere. Six to eight named beds may appear in the photograph of a single outcrop. The subsurface Lower Pennsylvanian, however, awaits further detailed study and differentiation. The lithologic changes under the Great Plains between the exposed sections of eastern Kansas and Nebraska and the western exposures in the Black Hills, eastern Wyoming, and Colorado are imperfectly known. Condra's zonation of the beds of the Hartville Uplift, Wyoming,[24] and its correla-

[16] Earle F. Taylor, M. W. Pullen, Paul K. Sims, and J. Norman Payne, "Methods of Subsurface Study of the Pennsylvanian Strata Encountered in Rotary-Drill Holes," *Ill. Geol. Surv., Rept. Investigations No. 93* (1944), pp. 9–21.

[17] "Pennsylvanian Stratigraphy near Grand Ledge, Michigan," *Jour. Geol.*, Vol. XLI (1933), pp. 79–86.

[18] "Traverse of Upper Des Moines and Lower Missouri Series from Jackson County, Missouri, to Appanoose County, Iowa," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XXV (1941), pp. 23–72.

[19] J. M. Weller, H. R. Wanless, L. M. Cline, and D. G. Stookey, "Interbasin Pennsylvanian Correlations, Illinois and Iowa," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XXVI (1942), pp. 1585–93.

[20] G. E. Condra and J. E. Upp, "The Middle River Traverse of Iowa," *Neb. Geol. Surv., Paper No. 4* (1933).

[21] "The Stratigraphy of the Pennsylvanian Series in Missouri," *Mo. Bur. Geol. and Mines*, Vol. XIII (2d ser., 1915).

[22] H. S. McQueen and F. C. Greene, "The Geology of Northwestern Missouri," *Mo. Geol. Surv. and Water Resources*, Vol. XXV (2d ser., 1938), pp. 120–24, 169–76.

[23] "The Stratigraphy of the Pennsylvanian System in Nebraska," *Neb. Geol. Surv., Bull. 1* (2d ser., 1927); "Correlation of the Pennsylvanian Beds in the Platte and Jones Point Sections of Nebraska," *Neb. Geol. Surv., Bull. 3* (2d ser., 1930).

[24] G. E. Condra and E. C. Reed, "The Permo-Pennsylvanian Section of the Hartville Area of Wyoming," *Neb. Geol. Surv., Paper 9* (1935).

tion with the mid-continent section is the most noteworthy such study up to the present.

The Pennsylvanian of Kansas is probably as well known as that of any area in the country.[25] Because of the presence there of both marine invertebrates and plant fossils well distributed throughout the column, Moore has proposed that the Kansas section be considered the standard Pennsylvanian succession of this continent, since the section in Pennsylvania does not afford a complete record of the succession of marine faunas.

In Oklahoma there is an abundance of information,[26] but the great changes in thickness and facies between northeastern Oklahoma[27] and the McAlester,[28] Ouachita,[29] Ardmore,[30] and Anadarko basins[31] will, no doubt, furnish grounds for research and difference of opinion for years to come. Dr. Cheney is contributing to the stratigraphy of the subsurface Pennsylvanian of Oklahoma and Texas basins in another paper in this symposium.

The Pennsylvanian of Arkansas varies from a thin northern Ozark facies to a tremendously thick southern Ouachita facies and offers problems similar to those of Oklahoma; but, with scantier marine faunas and less detailed study, these problems are less well understood.[32] There has been much difference of opinion as to the position of the Pennsylvanian-Mississippian boundary in Oklahoma and Arkansas, where the Jackfork, Stanley, and Springer formations have been variously interpreted.[33] Similar uncertainty exists with respect to the Parkwood of Alabama; the lower Pocahontas, Bluestone, and Princeton of West Virginia; some "transition" beds of the southern anthracite field; and the Amsden and perhaps other formations of the Rocky Mountains. It seems probable that along the southeastern and southern borders of the Appalachian and Ouachita geosynclines and in some parts of the Rocky Mountains deposition was nearly continuous from the Mississippian into the Pennsylvanian.

In Texas the outcropping Pennsylvanian of central and north-central Texas is well known from studies by F. B. Plummer and R. C. Moore[34] and by others. Dr. M. G. Cheney has worked

[25] R. C. Moore, "Stratigraphic Classification of Pennsylvanian Rocks of Kansas," *Kan. Univ. Bull. 22* (1935); John M. Jewett, "Classification of the Marmaton Group, Pennsylvanian, in Kansas," *Univ. Kan. Pub., Kan. Geol. Surv. Bull. 38*, Part XI (1941); and many other titles.

[26] Robert H. Dott, "Regional Stratigraphy of Mid-continent," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XXV (1941), pp. 1660–80.

[27] R. C. Moore, "Upper Carboniferous Rocks of Southeastern Kansas and Northeastern Oklahoma," *Kan. Geol. Soc., Guidebook, 11th Ann. Field Conf.* (1937), pp. 9–16.

[28] Thomas A. Hendricks, "Geology and Fuels Resources of the Southern Part of the Oklahoma Coal Field. I. The McAlester District, Pittsburg, Atoka, and Latimer Counties," *U.S. Geol. Surv. Bull. 874-A* (1937).

[29] H. D. Miser, "Carboniferous Rocks of the Ouachita Mountains," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XVIII (1934), pp. 971–1099.

[30] C. W. Tomlinson, "The Pennsylvanian System in the Ardmore Basin," *Okla. Geol. Surv. Bull. 46* (1929).

[31] W. B. Boyd, "Deepest Well in Mid-continent Region, Washita County, Oklahoma," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XXIV (1940), pp. 735–38.

[32] T. A. Hendricks and C. B. Read, "Correlations of Pennsylvanian Strata in Arkansas and Oklahoma Coal Fields," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XVIII (1934), pp. 1050–58.

[33] David White, "Age of Jackfork and Stanley Formations of Ouachita Geosyncline, Arkansas and Oklahoma, as Indicated by Plants," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XVIII (1934), pp. 1010–17; Chalmer L. Cooper, "Age Relations of Stanley and Jackfork Formations of Oklahoma and Arkansas," *Jour. Geol.*, Vol. LIII (1945), pp. 390–97.

[34] "Stratigraphy of the Pennsylvanian Formations of North Central Texas," *Univ. Tex. Bull. 2132* (1922).

extensively on the subsurface Pennsylvanian of the northern Texas area.[35] Both Dr. Plummer and Dr. Cheney are presenting recent findings on the northern Texas area in this symposium. Phillip King[36] and others have worked out the outcropping succession of the Marathon Basin in western Texas; and many areas of subsurface Pennsylvanian, especially in the Permian Basin, are known from oil test records of wells which have extended down to the Ordovician Arbuckle or Ellenburger limestones.

Little is known about the subsurface Pennsylvanian of the Dakotas. The outcropping succession of the Minnelusa in the Black Hills has been described and zoned.[37]

C. C. Branson,[38] A. K. Miller and H. D. Thomas,[39] H. W. Scott,[40] and, recently, the author have been interested in the Amsden, Tensleep, Casper, Quadrant, and other formations in the northern Rockies. Although a few good zone fossils have been found, a large part of the succession consists of barren quartzitic sandstones and nearly barren dense siliceous or dolomitic limestones. The boundaries between the major mid-continent divisions remain to be worked out. Fusulinid evidence shows that there are beds of Morrow, Lampasas, Des Moines, and Missouri age in Wyoming, and there may also be representatives of the Virgil.

In Colorado there are great changes in faciés between the thick, clastic, nonmarine Fountain formation of the Denver–Colorado Springs region and the finer-grained and more fossiliferous rocks of the Ingleside, Weber, Hermosa, McCoy, and other named formations in other parts of the state. L. G. Henbest[41] has recently summarized the succession in various parts of Colorado and correlated the zones with those of the mid-continent area. In some parts of the state fusulinids are useful aids in zonation and correlation. The position of the Pennsylvanian-Permian boundary in the thick Maroon formation is not yet determined, since the formation is almost wholly unfossiliferous. Recent work of J. Harlan Johnson,[42] R. Roth,[43] K. G. Brill,[44] and J. W. Vanderwilt[45] is noteworthy.

In New Mexico there are likewise great changes in thickness and facies between comparatively near-by areas. The rocks are, however, somewhat richer in

[35] "Geology of North Central Texas," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XXIV (1940), pp. 65–118.

[36] "The Geology of the Glass Mountains, Texas. I. Descriptive Geology," *Univ. Tex. Bull. 3038* (1930).

[37] Glenn S. Dillé, "Minnelusa of Black Hills of South Dakota," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XIV (1930), pp. 619–23.

[38] "Pennsylvanian Formations of Central Wyoming," *Bull. Geol. Soc. Amer.*, Vol. L (1939), pp. 1199–1226.

[39] "The Casper Formation (Pennsylvanian) of Wyoming and Its Cephalopod Fauna," *Jour. Paleon.*, Vol. X (1936), pp. 715–38.

[40] "Some Carboniferous Stratigraphy in Montana and Northwestern Wyoming," *Jour. Geol.*, Vol. XLIII (1935), pp. 1011–32.

[41] "Stratigraphy of the Pennsylvanian in the West Half of Colorado and in Adjacent Parts of New Mexico and Utah" (abstr.), *Bull. Amer. Assoc. Pet. Geol.*, Vol. XXX (1946), pp. 750–51.

[42] "Paleozoic Formations of the Mosquito Range, Colorado," *U.S. Geol. Surv. Prof. Paper 185B* (1934).

[43] "Type Section of Hermosa Formation, Colorado," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XVIII (1934), pp. 944–47.

[44] "Late Paleozoic Stratigraphy of West-central and Northwestern Colorado," *Bull. Geol. Soc. Amer.*, Vol. LV (1944), pp. 621–56; "Late Paleozoic Stratigraphy of Gore Area, Colorado," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XXVI (1942), pp. 1375–97.

[45] "Stratigraphy of Pennsylvanian Hermosa Formation in Elk Mountain, Gunnison County, Colorado," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XIX (1935), pp. 1668–77.

marine faunas than those in other Rocky Mountain areas. The Pennsylvanian has been recently studied by C. E. Needham[46] and zoned on the basis of fusulinids by M. L. Thompson,[47] and the results of correlation studies between several separate depositional basins by C. B. Read and G. H. Wood are presented in another paper in this symposium.

The thick Pennsylvanian successions of Idaho and Utah are imperfectly known. There is said to be at least a 10,000-foot section in the Beaverhead Range on the Idaho-Montana divide which awaits study. The Wells of southeastern Idaho and the Oquirrh of Utah have been studied, respectively, by J. S. Williams[48] and James Gilluly;[49] but both are very thick formations, and much remains unknown about them. The Pennsylvanian of the Uinta Mountains may exhibit lateral-facies gradation between the quartzitic sandstones of western Wyoming and Utah and the arkoses and marine limestones of Colorado, but such transition is inadequately described.

A part of the Supai formation of the Grand Canyon region may be of Pennsylvanian age, but it is largely barren of faunal remains, and the boundary between the Pennsylvanian and the Permian in the Supai has not been located. The Pennsylvanian of southern Arizona seems to be more nearly solid limestone than elsewhere in the United States. These rocks and their faunas await detailed study. In southeastern Utah the thick but local Paradox formation[50] contains the only important evaporites of the Pennsylvanian in the United States. They seem to record barred-basin accumulation.

The Pennsylvanian rocks of (1) Nevada, (2) southwestern California, (3) the Inyo Range of central California, (4) western Oregon, (5) the Blue Mountains of eastern Oregon, and (6) northern and northwestern Washington are poorly known as to age, lithology, succession, and fauna. Many are referred to simply as "Carboniferous" or "late Paleozoic." A small area in central Oregon has recently been carefully studied by C. B. Read and C. W. Merriam[51] and, on the basis of a fossil flora, has been referred to the early Pennsylvanian.

TECHNIQUES APPLIED TO PENNSYLVANIAN STUDIES

Newly advanced techniques applied to Pennsylvanian problems include the isolation of plant microfossils from coals and carbonaceous shales and the zonation of these rocks by a spore-succession column. Researches in this field by R. Kosanke and A. T. Cross are presented as part of this symposium. Other active students using this technique are J. M. Schopf, of the Bureau of Mines, Ray Bentall, and L. R. Wilson.[52]

[46] "Correlation of Pennsylvanian Rocks of New Mexico," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XXIV (1940), pp. 173–79.

[47] "Pennsylvanian System in New Mexico," *N.M. School of Mines, Bull. 17* (1942).

[48] "Carboniferous Formations in the Northern Blue Springs Hills, near Malad, Idaho," *Proc. Geol. Soc. Amer., 1935* (1936), p. 373.

[49] "Geology and Ore Deposits of the Stockton and Fairfield Quadrangles, Utah," *U.S. Geol. Surv. Prof. Paper 173* (1932), pp. 31–38.

[50] A. A. Baker, C. H. Dane, and J. B. Reeside, "Paradox Formation of Eastern Utah and Western Colorado," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XVII (1933), pp. 963–80.

[51] "A Pennsylvanian Flora from Central Oregon," *Amer. Jour. Sci.*, Vol. CCXXVIII (1940), pp. 107–11.

[52] J. M. Schopf, "Spores from the Herrin (No. 6) Coal Bed in Illinois," *Ill. Geol. Surv., Rept. Investigations No. 50* (1938); J. M. Schopf, L. R. Wilson, and Ray Bentall, "An Annotated Synopsis of Paleozoic Fossil Spores and the Definition of Generic Groups," *Ill. Geol. Surv., Rept. Investigations No. 90* (1944).

Although fusulinids have long been given a prominent place in Pennsylvanian studies because of their widespread distribution, complexity, and value as stratigraphic markers, other groups of microfossils—the ostracodes, conodonts, and smaller foraminifera—have received increasing attention in recent years from C. L. Cooper, who is participating in this symposium, Mrs. Helen Jeanne Plummer,[53] Mrs. Betty Kellett Nadeau,[54] and S. Ellison.[55]

Several groups of larger invertebrates, such as the single corals, trilobites, pelecypods, and Bryozoa, have been subjects of recent investigation by R. M. Jeffords,[56] J. M. Weller,[57] N. D. Newell,[58] R. C. Moore,[59] M. K. Elias,[60] J. B. Knight,[61] C. O. Dunbar and G. E. Condra,[62] and others. Weller summarizes the present and prospective usefulness of invertebrate fossils in interregional correlation and exact zonation of the Pennsylvanian in another paper in this program.

Leaf impressions have been relied upon for years as Pennsylvanian guide fossils, although many of the widespread and important species of plants are still manuscript species, awaiting adequate description and illustration. Charles Read, our principal student of Paleozoic leaf floras, presents the status of Pennsylvanian floras as aids in stratigraphic work in another paper in this symposium.

Electric well logging has permitted a greatly improved tracing of thin subsurface zones in the heterogeneous Pennsylvanian column as compared with the older drillers' logs or even samples furnished by the driller. Drilling-time logs, samples caught by the geologist at the well, and caliper surveys of hole diameter are other useful new tools in the subsurface Pennsylvanian. Radioactive logs have been used little thus far in stratigraphic work; but, because of greater concentration of radioactive materials in organic-rich zones, they may aid in picking out thin coal and carbonaceous shale beds.

Ecologic studies of Pennsylvanian faunas and floras have been made by comparatively few, notably David White,[63] M. K. Elias,[64] and a student now working under L. M. Cline on a statistical study of fossil frequency of vari-

[53] "Smaller Foraminifera in the Marble Falls, Smithwick, and Lower Strawn Strata around the Llano Uplift in Texas," *Univ. Tex. Pub. 4401* (1945), pp. 209–71.

[54] "Ostracodes of the Upper Pennsylvanian and Lower Permian of Kansas," *Jour. Paleon.*, Vol. VII (1933), pp. 69–108; Vol. VIII (1934), pp. 120–38; Vol. IX (1935), pp. 132–66.

[55] "Revision of the Pennsylvanian Conodonts," *Jour. Paleon.*, Vol. XV (1941), pp. 107–43.

[56] "Lophophyllid Corals from Lower Pennsylvanian Rocks of Kansas and Oklahoma," *Kan. Geol. Surv. Bull. 41* (1942), pp. 185–260.

[57] "Carboniferous Trilobite Genera," *Jour. Paleon.*, Vol. X (1936), pp. 704–14; "Evolutionary Tendencies in American Carboniferous Trilobites," *ibid.*, Vol. XI (1937), pp. 337–46.

[58] "Late Paleozoic Pelecypods: Pectinacea," *Kan. Geol. Surv.*, Vol. X (1937); "Late Paleozoic Pelecypods: Mytilacea," *ibid.*, Vol. X, Part II (1942).

[59] "New Crinoids from Upper Pennsylvanian and Lower Permian Rocks of Oklahoma, Kansas, and Nebraska," *Denison Univ. Bull.*, Vol. XXXIV (1939), pp. 171–279, and other crinoid articles.

[60] "Paleozoic Fenestrate Bryozoa," *Jour. Paleon.*, Vol. XI (1937), pp. 316–34.

[61] "The Gastropods of the St. Louis, Missouri, Pennsylvanian Outlier," *Jour. Paleon.*, Vol. IV (1930), suppl.; Vol. V (1931), pp. 1–15, 177–229; Vol. VI (1932), pp. 189–202; Vol. VII (1933), pp. 30–58, 359–92; Vol. VIII (1934), pp. 139–66, 433–47.

[62] "Brachiopoda of the Pennsylvanian System of Nebraska," *Neb. Geol. Surv. Bull. 5* (2d ser., 1932).

[63] "Climatic Implications of the Pennsylvanian Flora," *Ill. Geol. Surv. Bull. 60* (1931), pp. 271–82.

[64] "Depth of Deposition of the Big Blue (Late Paleozoic) Sediments in Kansas," *Bull. Geol. Soc. Amer.*, Vol. XLVIII (1937), pp. 403–32.

ous species in the Exline limestone of Iowa. Further studies of this sort will aid in understanding the significance of changes in lithology, both vertically and laterally, which affect the Pennsylvanian in nearly all areas of its outcrop.

Few studies have been made of insoluble residues of Pennsylvanian limestones, and these suggest that the technique, while useful, is somewhat less valuable than it has proved to be with the Cambro-Ordovician limestones and dolomites through studies by McQueen and others. Few mineralogical studies of Pennsylvanian sandstones have been made. Those few with whose results the author is familiar, principally in the Middle West, suggest that most of the particles are stable mineral species released from an older sediment and redeposited in the Pennsylvanian. Perhaps in the Rocky Mountains, where crystalline source rocks were closer at hand, such studies would throw more light on the provenance of the minerals. It is hoped that researches related to the exchangeable bases associated with clay minerals may aid in determining whether a particular clay or shale was deposited in fresh or marine waters.

Most paleogeographic maps have generalized the geography of a portion of the Pennsylvanian, not indicating that during such an interval of time the sea may have advanced eastward from Kansas to Illinois or Pennsylvania several times, with an equal number of periods of emergence. The accomplishment of more exact correlations, such as those of Cline and his associates, through the detailed tracing of beds from Kansas across Missouri and Iowa to Illinois should permit mapping the geography of a particular brief stage of one of the Pennsylvanian epochs to a degree heretofore never undertaken.

SUMMARY OF CLASSIFICATION OF THE PENNSYLVANIAN FORMATIONS OF TEXAS, WITH SPECIAL REFERENCE TO THE LOWER PENNSYLVANIAN OF THE LLANO REGION

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ABSTRACT

Middle and Upper Pennsylvanian beds of Texas are well known and are subdivided into a series of formations on the basis of lithology and fossils. The Lower Pennsylvanian, however, which outcrops in central Texas is more complex. Recent field work has distinguished three formations, one of which is subdivided into four members. These are briefly described and lists of characteristic fossils are presented.

GENERAL DESCRIPTION

The Pennsylvanian sediments of north-central Texas have a maximum thickness of about 5,400 feet. The lower 400–600 feet of this section consist of limestone; and the next 1,000–3,600 feet, of predominantly clastic sediments, are made up of siliceous and calcareous shales, thick sandstones, and a few limestones. The upper 1,600 feet consist of alternating limestones and calcareous shales, with a few coal seams, sandstone, and conglomerate layers. With the exception of groups of strata in the upper series, the Pennsylvanian formations in Texas do not show anything like complete cycles of sedimentation like those seen in the geologic section of Illinois, Kansas, and northern Oklahoma. The division of the strata into series, groups, and formations is made on a basis of lithology and paleontology. This system of classification, as far as the Upper Pennsylvanian strata go, is now quite satisfactorily established, and, except for minor details, Texas geologists are agreed fairly well on the nomenclature and correlations. On the other hand, the Lower Pennsylvanian formations, until recently, have been less well known paleontologically, and their divisions, classification, and correlation have been the subject of much discussion ever since the first work by Hill, Dumble, Cummins, and Sidney Paige.

CLASSIFICATION

The general classification of the strata in the middle and upper portions of the Pennsylvanian section in central Texas is shown in Table 1, to which has been added a list of the more common index fossils that characterize the various divisions.

The upper formations are classified into three groups: Strawn of Des Moines age, Canyon of Missouri age, and Cisco of Virgil age. These series are subdivided into convenient mappable units with a persistent, easily recognizable limestone at the top and, in most cases, a coarse sandstone or basal conglomerate at the base. In several cases an easily recognizable unconformity, with definite interformational channeling, separates the formations distinctly. The limestones have been mapped in great detail by oil geologists of at least three decades. In most cases they stand out clearly on aerial photographs and therefore form the most convenient basis for differentiation. On the other hand, with the exception of some interesting work by Wallace Lee, the intervening shales, sandstone,

and coals have been very largely neglected and deserve much more attention. A detailed study of northern Texas clastic sediments is a rich field for future research. Many geologists have asked why we do not drop the local Texas names—Strawn, Canyon, and Cisco—and adopt the wider-ranging mid-continent nomenclature, Des Moines, Missouri, and Virgil. This procedure in many cases can be adopted advantageously. On the other hand, because of the Arbuckle Mountains, the Cretaceous overlap, and change in facies along the Red River, it is not possible to trace any group of beds through from one area of deposition into another. Further, a very detailed study of fossil zones indicates that in many cases the successions of sediments in widely separated areas are not exactly equivalent—hiatuses occur in one area, expansion of sedimentation in another, transgressions of sediments in a third, etc. The use of a local name ties the particular formation or group under discussion to a given sedimentary area and indicates at once its geographical environment. This, in some

TABLE 1

CLASSIFICATION OF PENNSYLVANIAN FORMATIONS IN CENTRAL TEXAS

Series		Definition in Texas	Fusulinid Zones	Chonetid Zones	Ammonoid Zones
Texas	Mid-Continent				
Cisco.....	Virgil	Base of Cisco Lake ss. outcropping at Cisco L. spillway to top of Home Cr. lst. Includes lower Harpersville Fm. and the Thrifty and Graham Fms.	*Triticites ventricosa* (large robust forms) *Dunbarinella plummeri*	*Chonetes transversalis* *C. alata* *Lissochonetes plattsmouthensis*	*Uddenites serratus* *U. schucherti* *Shumardites simondsi*
Canyon...	Missouri	Top of Home Cr. lst. to base of L. Pinto ss. in north Texas and to base of Rochelle cgl. in central Texas. Includes Caddo Cr., Brad, Graford, and Lone Camp Fms.	*Triticites* (small forms) *T. ohioensis* *T. irregularis*	*Chonetina primitiva* *C. wyandottensis* *C. verneuilana*	*Shumardites cuyleri* *S. fornicatus* *Paraschistoceras reticulatum*
Strawn....	Des Moines	Base of Lake Pinto ss. or base of Rochelle cgl. to top Smithwick shale. Includes Mineral Wells (restricted), Garner, and Millsap Lake Fms.	*Fusulinella* (advanced type) *Fusulina* *Wedekindellina*	*Mesolobus euampyzus* *M. rochellensis* *M. mesolobus*	*Wellerites mohri* *Pintoceras postvenatum*
Bend.....	Atoka	Top of Smithwick shale to base of Big Saline lst. Includes Smithwick and Big Saline Fms.	*Fusulinella* (restricted type) *Profusulinella* *Pseudostaffella*	*Chonetes dominus* (large) *Paeckelmannia derelicta*	*Gastrioceras smithwickense* *G. occidentale* *Branneroceras branneri* *Phaneroceras compressum*
Sloan.....	Morrow	Base of Big Saline Fm. to top of Barnett shale or to top of Chappel or Ellenburger, where Barnett is absent. Includes Sloan Fm.	No fusulinids as yet observed in Texas	*Chonetes choteauensis*	*Phaneroceras nolinense*

cases, is a distinct advantage. On the other hand, system names and series names should be interstate and international to indicate long-distance correlations.

The Lower Pennsylvanian strata occur around the Llano region in central Texas (Fig. 1). These strata present more difficulties in classification than the upper ones do. Many attempts have been made, and many more will be made. The author's latest contribution is shown in Table 2. This differentiation of these strata into formations, like that of the upper portion of the Pennsylvanian section, is based on lithology and paleontology (Fig. 2).

DESCRIPTION OF UNITS

Sloan formation.—The Sloan formation is thin-bedded, cryptocrystalline, dense, black limestone. The fossils are numerous and fairly well preserved. The more typical forms are listed below:

LIST OF FOSSILS COLLECTED FROM THE SLOAN FORMATION

ANTHOZOA

Stereocorypha annectans Moore and Jeffords
Lophophyllidium adapterlum Moore and Jeffords
Barytichisma crassum Moore and Jeffords
B. callosum Moore and Jeffords
B. repletum Moore and Jeffords
Cladochonus fragilis Mather
Amplexocarinia corrugata (Mather)

CRINOIDEA

Ethelocrinus texasensis Moore and Plummer
Cibolocrinus punctatus Moore and Plummer
Diphuicrinus croneisi Moore and Plummer

BRACHIOPODA

Chonetes choteauensis Mather
Juresania wilberana (McChesney)
Paeckelmannia sp.
Horridonia globosa (Mather)
Marginifera roemeri Girty
Linoproductus nodosus (Newberry)
Dictyoclostus morrowensis (Mather)
Rhynchopora illinoisensis (Worthen)
R. magnicosta Mather
Spirifer opimus Hall
Punctospirifer campestris (White)
P. transversus (McChesney)
Composita ozarkana Mather
C. deflecta Mather
C. elongata Dunbar and Condra
Cleiothyridina orbicularis (McChesney)
Squamularia? perplexa (McChesney)
Hustedia miseri Mather

PELECYPODA

Edmondia, n.sp.
Myalina wyomingensis (Lea)
M. congeneris Walcott
Leda sp.
Parallelodon obsoletus Meek
Posidonia sp.
Pterinopectinella, n.sp.
Aviculopecten halensis Mather
Lithophaga sp.
Pleurophorus occidentalis Meek and Hayden
Allorisma reflexum Meek

GASTROPODA

Worthenia tabulata (Conrad)
Trepospira sp. indet.
Colpites cf. *C. minutus* (Sayre)
Naticopsis aff. *N. subnovata* Worthen
Orthonychia, n.sp.
Meekospira peracuta (Meek and Worthen)
Pharkidonotus, n.sp.

AMMONOIDEA

Phaneroceras nolinense (Cox)

On this list about 60 per cent of the species appear to have fairly close affinities with the Morrow of Arkansas. The other 40 per cent of the species appear to be slightly younger than the Morrow. The most characteristic fossils which are common and easily recognized are *Dictyoclostus morrowensis*, *Marginifera roemeri*, *Linoproductus nodosus*, and *Phaneroceras nolinense*. The Sloan may be recognized by this assemblage anywhere.

Big Saline formation.—The Big Saline formation is almost wholly limestone. It consists of six members—four east of

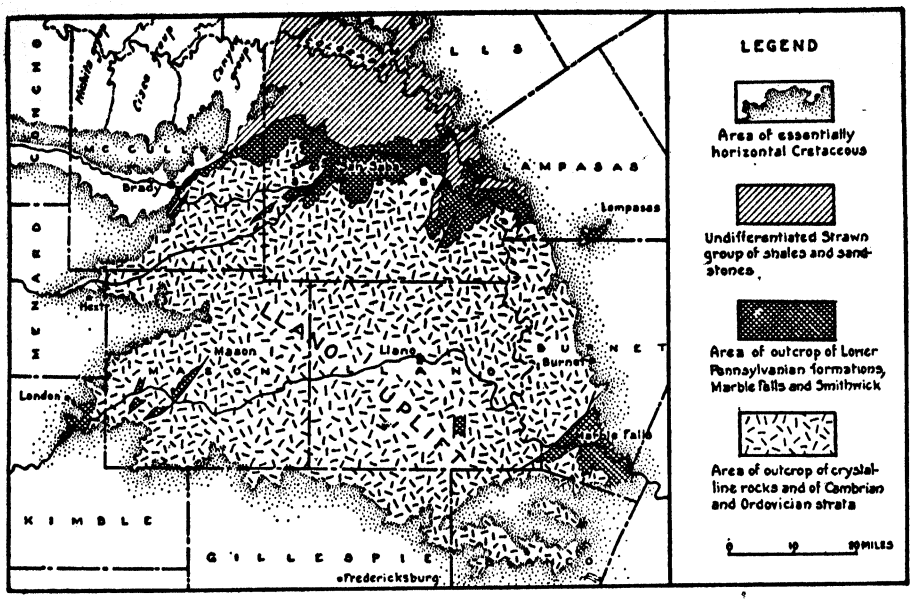

FIG. 1.—Map of the area around the Llano Uplift, showing the regional distribution and relationships of the older Paleozoics, the Lower Pennsylvanian, the Strawn group, Canyon group, and Cisco group of the Middle and Upper Pennsylvanian and the essentially horizontal Lower Cretaceous blanket that once covered the entire area but has been partially removed by the Colorado River and its tributaries. (Details of the Lower Pennsylvanian geology by F. B. Plummer; border of Lower Cretaceous in Gillespie County by V. E. Barnes; rest of map compiled from several published sources.)

TABLE 2

Group	Formation	Member	
		East of Cavern Ridge	West of Cavern Ridge
	Smithwick	Siltstone facies Shale facies	Absent .....
Bend (Atoka).....	Big Saline	Brister Lemons Bluff Aylor Gibbons	Soldiers Lemons* Bluff Brook Ranch Gibbons
Morrow (Upper Morrow).....	Sloan	Present	Absent

* Absent southwest of Brook Ranch in McCulloch County.

Cavern Ridge and two additional members west of the ridge, as shown in Table 2. A marked unconformity separates the Sloan formation from the Big Saline, and in many places the entire thickness of the Sloan was removed by erosion before the Big Saline strata were deposited. In the eastern area there is the Brister or massive limestone at the top, the Lemons Bluff sponge-spicule beds in the middle, the Aylor oölite below, and the Gibbons conglomerate at the base.

fossils are not plentiful and have not been studied in detail.

Lemons Bluff member.—The Lemons Bluff member of the Big Saline is quite distinctive. It is 20–60 feet thick and is made up of about 50 per cent sponge spicules and 50 per cent dense subcrystalline limestone. It occurs in bed 2–12 inches thick, separated by thin shale partings, so that the face of a cliff of this member looks like a masonry wall. It has been traced from Lampasas on

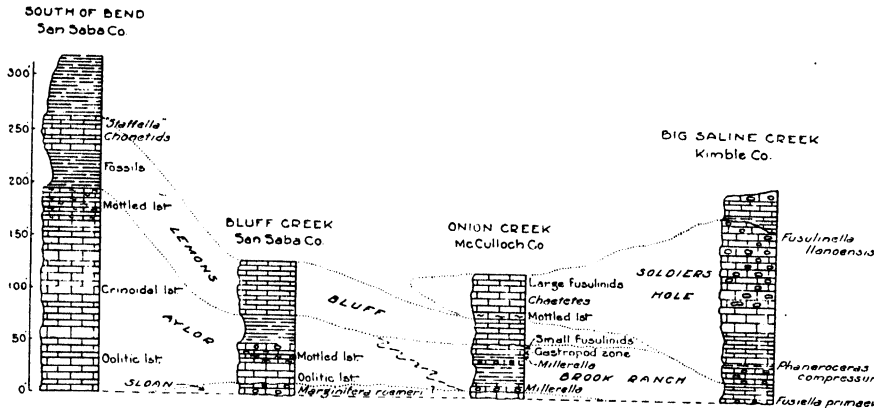

FIG. 2.—Stratigraphic sections, showing correlation of Pennsylvanian formations and members in the Llano region.

Gibbons conglomerate.—The lower or Gibbons conglomerate consists of a layer of chert and limestone pebbles, averaging ½ inch in diameter and 1 inch–3 feet thick, though commonly only about 6 inches thick. The layer is absent in many places, and in some cases the line of demarcation between the Big Saline and the Sloan is difficult to determine.

Aylor member.—The Aylor member has a bed of oölite 2–10 feet thick, which thickens toward the east and on the Colorado River has a total thickness of 100 feet, overlain by light-colored, coarsely crystalline, chert-bearing, massive, somewhat oölitic limestone. The

the east to Brady on the west, as shown in Figure 2, and its distinctive and easily recognizable characteristics make it of much aid in working out Lower Pennsylvanian stratigraphy. It may be identified easily by its spicules, which show up in thin sections and in insoluble residues, and by the fact that the high-silica content allows the rock to melt below 2000° F. The other Lower Pennsylvanian limestones do not melt but, upon heating, change to calcium oxide, with loss of carbon dioxide. The Lemons Bluff member of the Big Saline is quite fossiliferous. The following fossils have been identified:

MOST COMMON SPECIES FROM THE LEMONS BLUFF

FORAMINIFERA

Cribrostomum marblense H. J. Plummer
Staffella? sp.
Millerella marblensis Thompson

ANTHOZOA

Empodesma imulum Moore and Jeffords
Lophophyllidium extumidum Moore and Jeffords
L. idonium Moore and Jeffords
Paracaninia? sana Moore and Jeffords
Michelinia referta Moore and Jeffords
Striatopora religiosa Moore and Jeffords
Cladochonus texasensis Moore and Jeffords
Chaetetes favosus Moore and Jeffords

BRACHIOPODA

Neospirifer gorcii (Mather)
Spirifer opimum Hall
Rhipodomella pecosi (Marcou)
Chonetes dominus King
Chonetina sp.
Paeckelmannia derelicta King
Horridonia bullata (Mather)
H. globosa (Mather)

GASTROPODA

Pharkidonotus bendensis, n.sp.
Straparolus savagei Knight

AMMONOIDEA

Phaneroceras compressum (Hyatt)
Paralegoceras texanum (Shumard)
Branneroceras branneri (Smith)

ORTHOCERATOIDEA

Rayonnoceras croneisi, n.sp. (MS.)

NAUTILOIDEA

Ephippioceras divisum Hyatt
Coloceras sp.
Metacoceras sp.
Endolobus gibbosus Hyatt
Titanoceras sp.
T. illinoisensis (McChesney)
Solenochilus branneri Miller, Dunbar, and Condra

The most characteristic of these are *Neospirifer gorcii*, *Spirifer opimus*, *Chonetes dominus*, and *Paralegoceras texanum*.

Brister member.—The Brister, or upper, member consists of medium-thick beds of light-colored, coarsely crystalline limestone containing fossils quite similar to those in the Lemons Bluff member below, but generally richer in corals, especially *Chaetetes*, crinoids, and Bryozoa.

West of Cavern Ridge the strata are somewhat different in appearance and contain a different fauna, and therefore two of the members have been given different names from those on the east side. The Gibbons conglomerate consists of limestone pebbles, nodules, and water-worn fossils with almost no chert. The Lemons Bluff member thins out south of Brady, and the Soldiers Hole beds above rest directly on the Brook Ranch beds below.

Brook Ranch member.—The Brook Ranch beds are black, hard, subcrystalline, light-colored, cherty limestones and contain fossils quite different from those found in the Aylor member east of the ridge. Here are the primitive fusulinids, *Fusiella primaeva*, an interesting group of gastropods recently studied by Brookes Knight, and an entire absence of the oölitic deposits found in the Aylor. None of us has yet been able to correlate any of these Brook Ranch beds with strata east of Cavern Ridge.

Soldiers Hole member.—The upper, or Soldiers Hole, member is best developed southwest of Brady and along Big Saline Creek southeast of London. It consists of moderately massive, coarsely crystalline, light-colored limestone, in beds 2–12 feet thick, separated by distinct bedding planes. The most common fossils are as follows:

LIST OF FOSSILS FROM THE SOLDIERS HOLE MEMBER OF THE BIG SALINE FORMATION

FORAMINIFERA

Fusulinella llanoensis (Thomas)
Cribrostomum marblense H. J. Plummer
Endothyranella armstrongi sobrina H. J. Plummer
Endothyra distensa H. J. Plummer
Polytaxis scutella (Cushman and Waters)
Bradyina sp.

ANTHOZOA

Lophophyllidium extumidum Moore and Jeffords
Amplexocarinia corrugata (Mather)
Pseudozaphrentoides lepidus Moore and Jeffords
P. spatiosus Moore and Jeffords
Rodophyllum texanum Moore and Jeffords
Neokoninckophyllum simplex Moore and Jeffords
N. gracile Moore and Jeffords
Michelinia latebrosa Moore and Jeffords
Chaetetes eximius Moore and Jeffords
C. subtilis Moore and Jeffords
C. favosus Moore and Jeffords

BRACHIOPODA

Orthotetes robusta (Hall)
O. kaskaskiensis (McChesney)
Chonetes dominus King
Horridonia globosa (Mather)
Marginifera gallatinensis (Girty)
M. welleri Mather
M. haydenensis Girty
Linoproductus sp.
L. platyumbonus Dunbar and Condra
Dictyoclostus sp.?
Spirifer matheri Dunbar and Condra
S. opimus Hall
S. rockymontanus Marcou
Neospirifer cameratus (Morton)
N. goreii (Mather)
Punctospirifer transversus (McChesney)
Composita ozarkana Mather
C. elongata Dunbar and Condra
C. trilobata Dunbar and Condra
C. gibbosa Mather
C. wasatchensis Mather
Cleiothyridina orbicularis (McChesney)
Schisophoria resupinoides (Cox)
Roemerella sp.

PELECYPODA

Allorisma costatum Meek and Worthen
A. terminale Hall
Solenomya subradiata Herrick

GASTROPODA

Euphemites sp. indet.
Orthonychia, n.sp.
Mourlonia, n.sp.
Pharkidonotus, n.sp.

AMMONOIDEA

Pronorites llanoensis Plummer and Scott

The fossils indicate that the beds are younger than the Lemons Bluff strata, and they may prove to be equivalent to, or perhaps even younger than, the Smithwick. The most characteristic species are *Fusulinella llanoensis*, *Neospirifer cameratus*, *N. goreii*, *Dictyoclostus inflatus*, and *Pronorites llanoensis*.

Smithwick formation.—The Smithwick formation is black, fissile, highly siliceous shale, containing, in its upper portion, layers and lentils of fine, hard, firmly cemented siltstone. It is thickest in syncline areas and thins by loss of lower beds as it passes over anticlines, where its base, in some cases, is marked by a thin pebble conglomerate. Therefore, it appears that some deformation took place in central Texas at the end of the Big Saline epoch and that the Smithwick sediments filled the depressions and overlapped onto the elevations. There is a rather rich fauna in the basal Smithwick beds. The upper Smithwick in most places is unfossiliferous. The following list has been identified:

LIST OF FOSSILS COLLECTED FROM THE SMITHWICK FORMATION

ANTHOZOA

Hapsiphyllum tumidum Moore and Jeffords
H. retusum Moore and Jeffords
Zaphrentoides eccentricus Moore and Jeffords
Cumminsia aplata (Cummins)

Michelinia referta Moore and Jeffords
Palaeacis testata Moore and Jeffords
Zaphrentis sp.
Neokoninckophyllum arcuatum Moore and Jeffords

BRYOZOA

Fenestella cf. *F. regalis* Ulrich

BRACHIOPODA

Orbiculoidea missouriensis (Shumard)
Paeckelmannia derelicta King
Linoproductus sp.
Marginifera muricatina Dunbar and Condra
Dictyoclostus portlockianus (Norwood and Pratten)
Wellerella osagensis var. *immatura* Dunbar and Condra
Crurithyris planoconvexa (Shumard)
Cleiothyridina sp.
C. sublamellosa (Hall)
Hustedia brentwoodensis Mather
Juresania nebrascensis (Owen)

PELECYPODA

Solenomya sp.
Edmondia ovata (Meek and Worthen)
E. glabra Meek
Nucula, n.sp.
N. summa, n.sp.
Nuculopsis ventricosa (Hall)
Nuculana bellistriata (Stevens)
N. plicata, n.sp.
Nu. arata (Girty)
Conocardium snideri Morgan
Allorisma sp.

GASTROPODA

Pharkidonotus, n.sp.
Euphemites, n.sp.
Neilsonia, n. sp.
Glabrocingulum, n.sp.
Trepospira, n.sp.
Goniospira sp.
New g. and n.sp.
Straparolus savagei Knight
Amphiscapha subquadrata (Meek and Worthen)
Straparolus, n. sp.
Soleniscus (*Macrocheilina*) *ventricosus* (Hall)
New g. and n. sp.

CEPHALOPODA—PSEUDORTHOCERATIDEA

Orthoceras wapanuckense Girty
Pseudorthoceras knoxense (McChesney)
Mooreoceras normale Miller, Dunbar, and Condra
Bactrites postremus Miller
Dolorthoceras, n. spp.
Rayonnoceras hueconense Miller, Dunbar, and Condra

CEPHALOPODA—NAUTALOIDEA

Brachycycloceras normale Miller, Dunbar, and Condra
Coloceras sp.
Liroceras liratum (Girty)
Temnocheilus sp.
Metacoceras sp.
Solenochilus sp.

CEPHALOPODA—AMMONOIDEA

Pronorites sp.
P. arkansasensis Smith
Daraelites (*Boesites*) *scotti* Miller and Furnish
Nuculoceras smithwickense Plummer and Scott
Phaneroceras compressum (Hyatt)
Paralegoceras sp.
P. shumardi Plummer and Scott
Glaphyrites hyattianus (Girty)
G. angulatus (Girty)
G. raymondi Plummer and Scott
Goniolobocerás sp.
Gastrioceras occidentale (Miller and Faber)
G. smithwickense Plummer and Scott

The most characteristic fossils of the Smithwick are *Gastrioceras occidentale*, *G. smithwickense*, *Daraelites* (*Boesites*) *scotti*, *Cumminsia aplata*, and *Straparolus savagei*. None of these except the last one occurs in the Big Saline below.

Strawn-Smithwick contact.—Another deformational unconformity occurs at the top of the Smithwick. The Strawn strata were deposited on the deformed and somewhat eroded Smithwick beds. The unconformity evidently represents a long hiatus because not only is the Smithwick fauna in every way unlike the Strawn fauna, but much compaction settling and widespread deformation took place before the Strawn strata were deposited. In many places the deformation was such that the Smithwick and Marble Falls strata dip east while the Strawn beds are inclined to the west. In

fact, the unconformity between the Bend and the Strawn is the most pronounced in a section replete with unconformities.

SUMMARY

It will be apparent from these brief descriptions that the Lower Pennsylvanian strata of central Texas, long divided into only two formations—the Marble Falls limestone below and the Smithwick shale above—are now divisible into three formations—Sloan, Big Saline, and Smithwick—recognizable on a basis of lithology and paleontology. The Big Saline formation, by its fossils, pebble conglomerate, and sponge-spicule content, is easily separated into four members on each side of the Cavern Ridge barrier. Furthermore, the beds on the west side of the ridge are apparently somewhat younger than stratigraphically equivalent strata on the west, indicating that the Big Saline sea in central Texas transgressed from the east toward the west, so that slightly younger and more advanced types of organisms differentiate the beds in the two regions. For this reason two of the members—the Soldiers Hole and the Brook Ranch—on the west side of the ridge are given different names from beds of equivalent stratigraphic position on the east side. The diagram (Fig. 2) illustrating typical geologic sections from east to west across the Llano region will perhaps make these relationships somewhat clearer.

PENNSYLVANIAN CLASSIFICATION AND CORRELATION PROBLEMS IN NORTH-CENTRAL TEXAS

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ABSTRACT

This paper is intended as a commentary on past and present classifications and correlations of the Pennsylvanian rocks of north-central Texas, with suggestions regarding solution of some of the regional problems involved. It is apparent that the problems are too complex and numerous to expect early settlement. The following major premises and conclusions are presented:

1. The Pennsylvanian rocks of north-central Texas may now be correlated with considerable certainty with those of other regions.
2. A time-stratigraphic classification in agreement with a standard time-stratigraphic classification for the system would have important advantages.
3. The principal standard section of reference should be established in the northern Appalachian region, with time-stratigraphic divisions based, in part, on data available from other regions, particularly the Mid-Continent.
4. The standard and regional time-stratigraphic divisions should be based on broad studies of structural and lithologic data, as well as on presently available paleontological evidence, in the hope of establishing widely recognizable, distinctive, and equivalent time-stratigraphic divisions which would require a minimum revision of boundaries heretofore used in the various major Pennsylvanian provinces.
5. Division of the Pennsylvanian system into six series appears appropriate, the major Pottsville and Pittsburgh divisions of the northern Appalachian region each being divided into three series, known, respectively, in north-central Texas as the Springer, Morrow, and Lampasas and the Strawn, Canyon, and Cisco.
6. The highest position used for the boundary between these lower and upper major divisions in Pennsylvania appears to agree closely with the usage of the last three decades in north-central Texas, where the boundary between the "subsurface Bend" and the Strawn has been placed above the "Caddo" limestone of oil-field terminology.
7. The top of the Caddo limestone occurs at a variable position in this region. The upper boundary of the youngest Caddo limestone has been used herein as a boundary between the Lampasas and the Strawn series. It is thought to be approximately equivalent to the Dornick Hills–Deese boundary of the Ardmore basin, the Boggy-Thurman boundary of eastern Oklahoma, the Tradewater-Carbondale boundary of the *Illinois Geological Survey Bull.* 67 (1942), the long-established Kanawha-Allegheny boundary of West Virginia, and the boundary recommended in Pennsylvania on the basis of recent studies. A somewhat lower boundary may ultimately prove more useful.

INTRODUCTION

Extensive drilling for petroleum in north-central Texas has revealed that the Pennsylvanian rocks of this region have a composite thickness of about 10,000 feet as measured in wells from the Fort Worth Basin westward. This is 5,000 feet greater than the maximum composite outcrop section and nearly 9,000 feet thicker than the Pennsylvanian section beneath Permian rocks on the buried major arches in the north and southwest parts of this region, namely, the Muenster and Electra arches of the Red River Uplift and the Concho arch, which extends north-northwest from the present Llano Uplift. Obviously, many problems confront the effort to classify and correlate such a great volume of sedimentary rocks. Complications arise not only from the numerous lithologic, structural, and biotic variables but from the extensiveness of the region and the inadequacy of data in areas of limited drilling or where the Pennsylvanian rocks are covered by the Comanche overlap. Much additional geological information is, of course, expected from future drilling, stimulated by past oil discoveries which total no less than 2,000,000,000 barrels in this region.

Aided in large part by fusulinid

studies, correlations from surface to subsurface are becoming more firmly established in north-central Texas. Transition from classification by local lithologic groups to time-stratigraphic series is progressing, although along somewhat divergent lines, according to the varying degree of emphasis given to (1) paleontologic, structural, and stratigraphic criteria; (2) surface versus subsurface usage; and (3) the classification to which the individual geologist is accustomed, chiefly north-central Texas versus northern Mid-Continent usage.

PROBLEMS OF CLASSIFICATION

GROUPS VERSUS TIME-STRATIGRAPHIC SERIES

Initial studies of Pennsylvanian outcrops in the Brazos and Colorado River valleys of Texas by Cummins, Tarr, Drake, and others a half-century ago gave excellent descriptions of the rock sequences and logical designation of certain divisions or lithologic groups, based largely on similarity of rock characteristics. Similar work progressing in other states resulted in the development of many local classifications, each with names locally derived. The expansion of geological studies has been hampered by the resultant dissimilarity of division boundaries and diversity of names.

The need of time-stratigraphic divisions has become increasingly evident. Knowledge of time relationships of the principal geological features within each basin and the interrelationships between the several major basins has both scientific and economic value. Stratigraphical, structural, and paleontological studies of Pennsylvanian rocks are of special interest because of the enormous energy resources found therein. The 1927 and 1935 Heerlen conferences on the Carboniferous reflect the well-recognized international importance of that part of the geologic column which is commonly called the "Pennsylvanian" and "Mississippian" systems in the United States.

During Pennsylvanian time, downward crustal movements in the Appalachian, southern Mid-Continent, and Marathon regions permitted deposition of clastic shallow-water sediments, several miles thick. By the close of Pennsylvanian time these southwest-trending troughs had become belts of extreme folding, uplift, and thrust faulting. The main axes of crustal adjustments migrated frontally to some extent during successive stages of subsidence, folding, thrusting, elevation, and erosion. To the northwest of these belts of extreme crustal adjustments there developed a number of broad domes, arches, flexures, and intervening basins, more or less transverse but doubtless influenced in large degree by forces and events in the more dominant belt of geosynclines and mountain-building. Extensive foreland areas remained near sea level, as revealed by innumerable alternations from marine to nonmarine deposits. The resulting great lateral changes in the thickness and character of the sediments and in the ecological conditions and subsequent succession of structural modifications have all contributed to the complex whole and have multiplied the difficulties of correlations and of establishment across the continent of equivalent time-stratigraphic divisions of this great rock system.

APPALACHIAN AND MID-CONTINENT STANDARD SECTIONS

Data from the various Pennsylvanian provinces in North America give increasing support to the view that, within these several related troughs, basins, and foreland areas, major changes took place

with sufficient contemporaneity that significant partition of Pennsylvanian time into several major epochs can be made with considerable reliance. That certain coal and marine limestones can be traced for hundreds of miles has been pointed out by many writers. The observed cyclic deposition and marked floral changes imply important fluctuations of sea level and climatic changes, which would leave very widespread, nearly simultaneous, records in the swamp, playa, and shallow marine deposits of the Pennsylvanian system. It therefore appears both desirable and feasible to work toward the establishment of equivalent time-stratigraphic divisions of this system in the various Pennsylvanian provinces. A major step in this direction was the publication in 1944 of the report on correlation of Pennsylvanian formations of North America by a National Research Council subcommittee, R. C. Moore, chairman.[1] Many specialists in Pennsylvanian stratigraphy from various parts of the United States aided in the preparation of this correlation chart, with its five major time-stratigraphic divisions and its more than sixty regional sections. A subsequent report[2] by a subcommittee of the Geologic Names and Correlation Committee of the American Association of Petroleum Geologists advocated worldwide recognition of the Pennsylvanian rocks as a system rather than as a series or subsystem and the establishment in North America of costandard Appalachian and Mid-Continent standard sections having six main time-stratigraphic divisions. In the latter report the Lower Pennsylvanian was divided into two series, instead of one as in the National Research Council subcommittee report. A forthcoming paper will review certain evidence supporting division of the Lower Pennsylvanian of the Mid-Continent region into two series (Springer and Morrow). More than one hundred and twenty-five species of invertebrate fossils have now been collected and identified from the 4,000 feet of Springer rocks in the Ardmore Basin by G. E. Condra and M. K. Elias. The assemblage is clearly intermediate between Chester and Morrow faunas, according to these investigators.[3]

Classification by standard time-stratigraphic divisions, differentiated by major biotic, structural, and depositional changes, should eventually bring much greater stability and usefulness to the various provincial stratigraphic classifications than would be attainable otherwise.

PRINCIPAL PENNSYLVANIAN STANDARD REFERENCE SECTION

Use of the Appalachian and Mid-Continent sections of the Pennsylvanian system as costandards gives recognition to the equal, or nearly equal, importance, as well as to the complementary nature, of these two sections for purposes of Pennsylvanian classification. Some geologists have advocated that the Mid-Continent Pennsylvanian should be used as the principal standard because of better development in the upper part of the system, excellent outcrops, central location, greater marine content, more diverse structural history, and greater progress in classification of these rocks into time-stratigraphic divisions. Others

[1] R. C. Moore *et al.*, "Correlation of Pennsylvanian Formations of North America," *Bull. Geol. Soc. Amer.*, Vol. LV (1944), pp. 657–706.

[2] M. G. Cheney *et al.*, "Classification of Mississippian and Pennsylvanian Rocks of North America," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XXIX (1945), pp. 125–69.

[3] Personal communications, March and December, 1946.

maintain that the principal standard section should be in Pennsylvania and vicinity, where the type section of the system is located and where differentiation into series first began, the name "Allegheny series" having been proposed by Rogers in 1840.

Plant fossils are more abundant and better known in the Appalachian region than in the Mid-Continent. Doubtless, useful nonmarine faunas are also present as in the western European Coal Measures.[4] Several marine invasions extended into the Appalachian region during Upper Pottsville and subsequent Pennsylvanian time, leaving representative fusulinid, brachiopod, and other faunas. Because coal deposits are of such great economic importance and are particularly characteristic of the Pennsylvanian system, it seems appropriate that the type and principal standard section should be in an area having numerous and widespread coal beds and abundant, well-preserved coal-forming floras, as is especially true of the Appalachian region.

Most geologists in the United States prefer to use the name "Pennsylvanian system" rather than "Upper Carboniferous series." This implies the use of the northern Appalachian section as a world type and standard for this system. The Appalachian section is both closer and more similar to the western European Upper Carboniferous or Coal Measures than is the Mid-Continent section; therefore, it is probably more useful for correlation of the North American and European sections.

Contrary to earlier impressions, accumulating data now indicate that the flora and fauna are in virtual agreement in the correlation of much of the Appalachian, Illinois, and Mid-Continent sections. The paleontological data and relationships have been discussed at some length in the two subcommittee reports previously mentioned. Present regional and interregional co-ordination awaits as much or more upon solution of problems of classification as of correlation. The choice of a principal section of reference and its organization into major time-stratigraphic divisions appears essential to the development of a more complete understanding of Pennsylvanian rocks of North America and of any portion thereof.

A questionnaire sent to 28 state geological surveys in 1941, requesting a statement of preference as to the most desirable standard section of reference for the Pennsylvanian system brought 22 replies; 15 favored the Pennsylvania, West Virginia, Ohio region, 5 the Mid-Continent, and 1 each West Texas and New Mexico.

NAMES OF STANDARD MID-CONTINENT SERIES

The foregoing considerations encourage efforts to classify the Pennsylvanian rocks of north-central Texas and elsewhere into time-stratigraphic or series divisions to replace earlier rock groups. The rules of nomenclature provide for revision of boundaries and change of rank of a stratigraphic sequence without change of name. Well-established provincial names are not readily displaced, and, furthermore, they have both geographical and more or less definite time significance. An attempt to revise the north-central Texas Pennsylvanian into divisions having series rank was made in 1939 as part of a symposium on the geology of West Texas.[5] The changes then

[4] A. E. Trueman, "Stratigraphical Problems in the Coal Measures of Europe and North America," *Quart. Jour. Geol. Soc. London*, Vol. CII (1946), pp. xlix–xcii and Pl. A.

[5] M. G. Cheney, "Geology of North-central Texas," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XXIV (1940), Fig. 9 and p. 109.

proposed have found acceptance in part, but no concordance of usage has as yet become established. Some geologists coming to Texas from the northern Mid-Continent region have advocated the replacement of Texas names by those of the related, though separate, province north of the Arbuckle Mountains. Such a proposal was made in 1946 in a stimulating paper by R. C. Spivey and T. G. Roberts.[6] Texas geologists, interviewed in 1939 on this question, expressed an overwhelming preference for the adoption of well-known Texas names for series to be set up in north-central Texas, such series to have boundaries equivalent to those established or appropriate for the northern Mid-Continent region.

Use of the Springer, Morrow, Lampasas, Des Moines, Missouri, and Virgil as the standard Mid-Continent series sequence appears to be proper, based on the rule of priority. It may prove desirable to use each of these names only in the particular province of its origin. Elsewhere correlative divisions may bear appropriate provincial series names and be referred to as of Springeran, Morrowan, Lampasan, Desmoinesian, Missourian, or Virgilian age to show time relationships. Suffixes have been used sparingly in this paper, although, admittedly, exactness of meaning and greater consistency are achieved by the use of suffixes such as *-ian*, *-an*, *-ous*, *-ic*, or *-ene* to indicate time and time-stratigraphic significance. On the other hand, designation of rank, whether series, group, or formation, shows the meaning intended. Brevity and euphony are more commonly retained and redundancy avoided when the suffix is omitted. Opposing views on this subject have been ably reviewed by C. W. Tomlinson.[7]

NORTH-CENTRAL TEXAS STRATIGRAPHY

The Pennsylvanian of north-central Texas presents numerous difficulties of classification and correlation, chiefly because of incomplete paleontological data, lack of index fossils in parts of the section, changing facies, wide separation between some outcrop areas, and great differences between basin and uplift sections.

Chart I shows the classification and terminology of this and certain earlier papers, particularly *University of Texas Bulletin 3232*.[8] Areas where five of the six Pennsylvanian series crop out in north-central Texas are shown in Figure 1. The Springer series is represented in wells in the northeast part of this region but has not been recognized at the surface.

NEW NAMES AND NEW USAGE

Since original publication of Chart I seven years ago, accumulating geological data appear to necessitate certain changes in the classification and correlation of the lower part of the Pennsylvanian section. As reported in 1945[9] fusulinids in cores from the uppermost "subsurface Bend" of Coleman County are similar to those found in the outcropping Dennis Bridge and Kickapoo Falls limestones of southwestern Parker and northwestern Hood counties. They indicate approximate time equivalence of these beds with upper Dornick Hills and Savanna-Boggy divisions of Oklahoma,

[6] "Lower Pennsylvanian Terminology in Central Texas," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XXX (1946), pp. 181–86.

[7] "Technique of Stratigraphic Nomenclature," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XXIV (1940), pp. 2038–48.

[8] E. H. Sellards, "The Geology of Texas. Vol. I," *Univ. Tex. Bull. 3232* (1932), pp. 98–115.

[9] Pp. 162–63 of ftn. 2.

CHART I

UNIV. TEXAS BULL. 3232			M.G. CHENEY, A.A.P.G. BULL., VOL. 24, NO. 1 (1940) VOL. 29, NO. 2. (1945)				
	GROUP	FORMATION	SERIES	GROUP	FORMATION (EQUIVALENT)	IMPORTANT MEMBERS	
PERMIAN	DOUBLE MOUNTAIN	WHITEHORSE	GUADALUPE	WHITEHORSE			PERMIAN
		BLAINE		EL RENO (SAN ANDRES)	DOG CREEK SH BLAINE GYP FLOWER-POT SH.		
		SAN ANGELO			SAN ANGELO SS		
	CLEAR FORK	CHOZA VALE	LEONARD	CLEAR FORK	CHOZA VALE	MERKEL DOL BULLWAGON DOL	
		ARROYO			ARROYO	STANDPIPE LS. LYTLE LS. RAINY LS.	
	WICHITA	LUEDERS		LUEDERS	LAKE KEMP MAYBELLE LS. PAINT ROCK LS		
		CLYDE		CLYDE	TALPA GRAPE CREEK		
		BELLE PLAINS		BELLE PLAINS x	BEAD MOUNTAIN VALERA SH & ANHYDRITE + JAGGER BEND LS ELM CREEK LS JIM NED SH +	BEAVERBURK	
		ADMIRAL	WOLFCAMP	ADMIRAL x	FISK + HORDS CREEK x		
		PUTNAM		PUTNAM	COLEMAN JUNCTION x SANTA ANNA BRANCH x		
		MORAN		MORAN	SEDWICK x "HORSE CREEK" x "DOTHAN" x		
PENNSYLVANIAN	CISCO	PUEBLO		PUEBLO x	CAMP COLORADO x STOCKWETHER x		
		HARPERSVILLE			SADDLE CREEK x	COON MOUNTAIN SS WALDRIP LS. NO. 1 " " " NO. 2	
		THRIFTY	CISCO x	THRIFTY x	OBREGON + CHAFFIN x (CRYSTAL FALLS) BRECKENRIDGE x SPECK MOUNTAIN (BLACH RANCH) x IVAN (ROCKY MOUND, BELLEROPHON) x AVIS SS	WALDRIP-NEWCASTLE COALS PARKS MOUNTAIN SS SS	UPPER
		GRAHAM		GRAHAM	WAYLAND SH GUNSIGHT LS (CYCLE 9?) POST-BUNGER CYCLES 1-7 x BUNGER x NORTH LEON x		
					GONZALES x	"UPPER LS." SALEM SCHOOL LS KISINGER CHANNEL	
	CANYON	CADDO CREEK	CANYON x	CADDO CREEK	HOME CREEK LS HOG CREEK SH		
		BRAD		BRAD	RANGER LS PLACID SH (SEAMAN RANCH)		
		GRAFORD		GRAFORD	WINCHELL CEDARTON SH (WOLF MOUNTAIN) ADAMS BRANCH	"LIMESTONES 1-4" "FUSULINID BED" DALTON COAL "FUSULINID BED"	
		PALO PINTO		WHITT +	PALO PINTO x (BROWNWOOD SH)	WILES LS. POSIDEON SH. WYNN LS. +	
	STRAWN	MINERAL WELLS			KEECHI CREEK x (BROWNWOOD SH)	---------SS ---------LS ---------LS	
					SALESVILLE x (BROWNWOOD SH)	TURKEY CREEK SS ---------SS ---------LS ---------SS	
			STRAWN x	LONE CAMP +	EAST MOUNTAIN SH	DOG BEND LS LAKE PINTO SS VILLAGE BEND LS (U PART CAPPS LS)	MIDDLE
		GARNER			GARNER	HOG MOUNTAIN SS BRAZOS R. RICKER CO. MINGUS SH & SS	
		MILLSAP LAKE		MILLSAP LAKE x x	GRINDSTONE CREEK LAZY BEND x x	THURBER COAL GOEN LS. RICKER STA. SANTO LS. LS "BUCK CREEK SS" BRANNON BRIDGE LS HILL CREEK	
	BEND	SMITHWICK	LAMPASAS +	KICKAPOO CREEK ++	RAYVILLE ++ PARKS + CADDO POOL + (CADDO LIME, BEND SUBSURFACE)	DENNIS BRIDGE KICKAPOO FALLS DICKERSON	
				BEND (ATOKA EQUIV)	SMITHWICK	LAKE SS PAY	
		MARBLE FALLS	MORROW		BIG SALINE x (MARBLE FALLS)	"RANGER" PAY McCLESKY SS PAY	LOWER
MISS.	CHESTER OSAGE	BARNETT CHAPPEL	CHESTER-MERAMEC OSAGE		SLOAN +++ BARNETT CHAPPEL		MISS.
ORD.	ELLENBURGER	COT-JEFF CITY (EQUIVALENT)		ELLENBURGER	COTTER-JEFF CITY-HONEYCUT * (EQUIV.)		ORD.
		ROUBIDOUX			ROUBIDOUX GORMAN TANYARD	STAENDEBACH DOL LS THORNHILL LS DOL *	
		GASCONADE			GASCONADE		
		EMINENCE (EQUIVALENT)			EMINENCE		
		POTOSI			POTOSI * WILBERNS	PEDERNALES DOL *	
CAMBRIAN		SIGNAL MT.			SAN SABA	SAN SABA LS.	CAMBRIAN
		FORT SILL			WILBERNS	POINT PEAK SH. MORGAN CK. LS.	
		WILBERNS			CAP MOUNTAIN RILEY *	WELGE SS.	
		CAP MT.			HICKORY	LION MT. SS. CAP MT. LS. HICKORY SS	
		HICKORY					

+ NEW NAMES X REDEFINED ——CHANNEL AND SANDSTONE DEPOSITS (BULL A.A.P.G. VOL 24 NO 1)

* CAMBRIAN ORDOVICIAN OF CLOUD, BARNES AND BRIDGE JUNE 1945 UNIV. TEX. PUB 4301

++ REVISED DEC. 1946, MGC. +++ F.B. PLUMMER UNIV. TEX. PUB. 4401

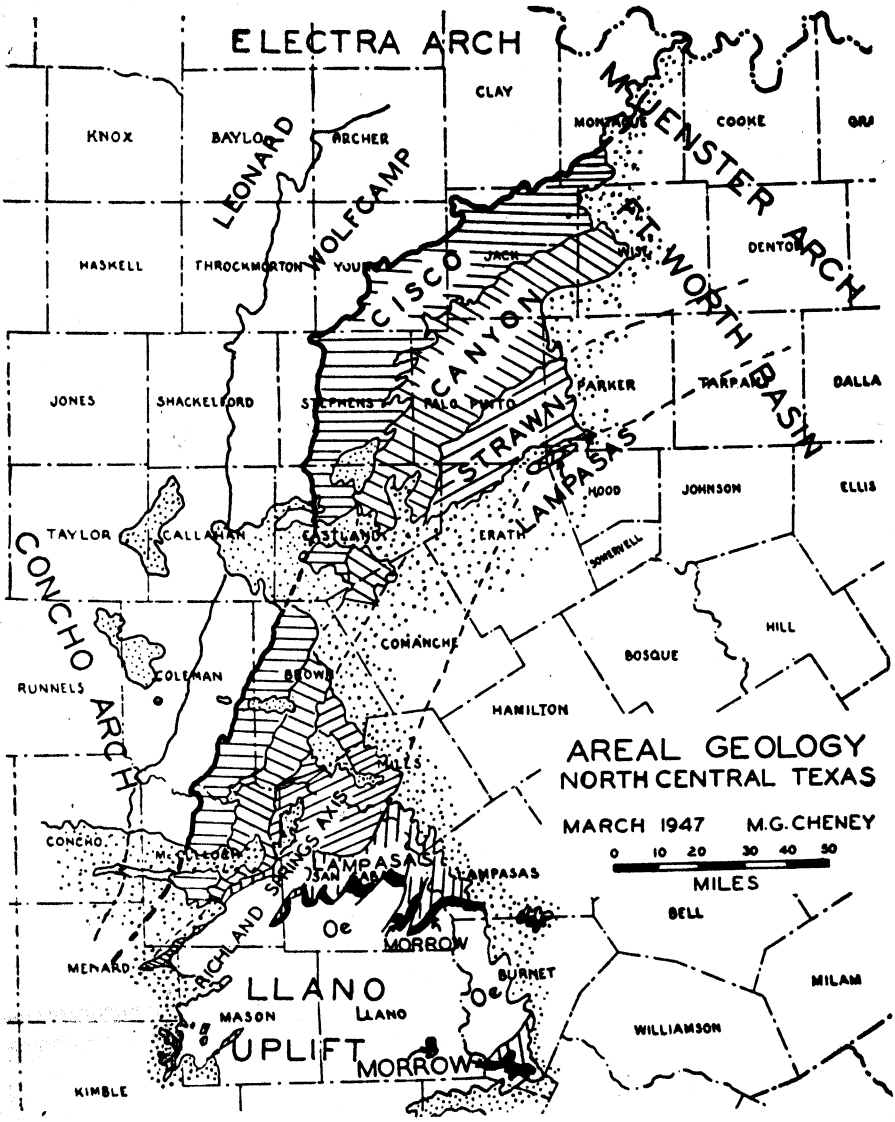
ELECTRA ARCH
KNOX
BAYLOR
ARCHER
CLAY
MONTAGUE
COOKE
GRA
LEONARD
WOLFCAMP
MUENSTER ARCH
FT. WORTH BASIN
HASKELL
THROCKMORTON
YOUNG
JACK
WISE
DENTON
CISCO
CANYON
JONES
SHACKELFORD
STEPHENS
PALO PINTO
PARKER
TARRANT
DALLA
STRAWN
LAMPASAS
HOOD
JOHNSON
ELLIS
TAYLOR
CALLAHAN
EASTLAND
ERATH
SOMERVELL
CONCHO ARCH
COMANCHE
HILL
BOSQUE
RUNNELS
COLEMAN
BROWN
HAMILTON
HILLS
AREAL GEOLOGY
NORTH CENTRAL TEXAS
MARCH 1947
M.G. CHENEY
0 10 20 30 40 50
MILES
CONCHO
McCULLOCH
LAMPASAS
SAN SABA
LAMPASAS
BELL
RICHLAND SPRINGS
Oe
MORROW
MENARD
BURNET
MILAM
LLANO
MASON
LLANO
UPLIFT
MORROW
WILLIAMSON
KIMBLE

FIG. 1

which, in turn, according to evidence of the fusulinids, brachiopods, and flora, are close time equivalents of the Upper Tradewater of Illinois and the Clarion sequence of eastern Ohio and western Pennsylvania.[10] Clarion beds have been classified as basal Allegheny by some, but as Upper Pottsville in more recent reports by G. H. Ashley, state geologist of Pennsylvania.[11] These considerations favor placing the Lampasas-Strawn series boundary at the disconformity between the Dennis Bridge limestone and the overlying conglomerate. When thus restricted, the Strawn series of the Brazos River section is about 2,000 feet thick, its two groups and four formations measuring approximately as follows:

Strawn Series	Feet
Lone Camp group:	
East Mountain shale formation	400
Garner formation	200
Millsap Lake group (restricted):	
Grindstone Creek formation	650
Lazy Bend formation (restricted)	750

A "Kickapoo Creek group" is now proposed for the section between Millsap Lake group (restricted) and the Bend group, in other words, from the disconformity at or near the top of the Dennis Bridge limestone down to the top of the Smithwick. This new name is derived from Kickapoo Creek of southwestern Parker and northwestern Hood counties, Texas. The upper 200 feet, more or less, of this group outcrops between the villages of Dennis and Lipan. However, the Kickapoo Creek group must be studied mostly from well samples. Data from wells indicate that approximately 2,600 feet of the section should be assigned to this group in southwestern Parker County, 3,500 feet or more in the Fort Worth Basin.

Logs of the Gilbert No. 1 and Wheeler No. 1 wells of southwestern Parker County, 5 miles north-nourtheast of Lipan, were shown in a cross section published in 1945[12] and are here reproduced in modified outline as Figure 2.

The stratigraphic section assigned to the Kickapoo Creek group includes the Rayville (new name defined below), Parks, and Caddo Pool formations, which extend from depths of approximately 380–1,350, 1,350–2,125, and 2,125–3,000 feet, respectively, in these wells. It seems appropriate to refer to the Kickapoo Creek group as "Upper Lampasas" and the Smithwick and Big Saline formations as "Lower Lampasas."

As tentatively correlated with the more clastic Colorado River outcrop section, 100 miles to the southwest, the Kickapoo Creek group includes Drake's Lynch Creek beds (No. 4) up to the top of the thick, "blackish" shales of the Big Valley beds (No. 12), a total thickness of about 2,000 feet.[13] The equivalent in the McAlester Basin of Oklahoma is thought to extend from the top of the Atoka to the top of the Boggy, with maximum thickness of about 7,000 feet. The Kickapoo Creek group is now proposed in preference to the elevation of Parks formation to Parks group, as suggested in 1945.[14] Evidently, the uppermost part of the Caddo limestone in Coleman and Brown counties is somewhat younger than the limestones and shales heretofore classified as the Parks formation in Young and Stephens coun-

[10] *Ibid.*, pp. 146–56.

[11] "The Pittsburgh-Pottsville Boundary," *Jour. Geol.*, Vol. LIII (1945), pp. 374–89; see also ftn. 1 above.

[12] Fig. 2, pp. 154–55 of ftn. 1.

[13] N. F. Drake, "Report on the Colorado Coal Field of Texas," *Tex. Geol. Surv., Fourth Ann. Rept.* (1893), reprinted as *Univ. Tex. Bull. 1755* (1917), pp. 17–21.

[14] Pp. 163–64 of ftn. 1.

ties. The Caddo Pool formation is apparently assignable to the Kickapoo Creek group, according to evidence reported by Spivey and Roberts.[15]

The name "Rayville formation" is here proposed for beds from the disconformity just above the Dennis Bridge limestone down to the top of the Parks formation. It includes a total thickness of approximately 1,000 feet of alternating limestone, blue and black shales, and sandstones in the type locality of southwestern Parker County. The name "Rayville" is derived from an early settlement, now almost abandoned, located 3.4 miles southwest of Dennis and a like distance northeast of Kickapoo Falls. Important outcropping members include the Dennis Bridge limestone; an un-

[15] P. 183 of ftn. 6.

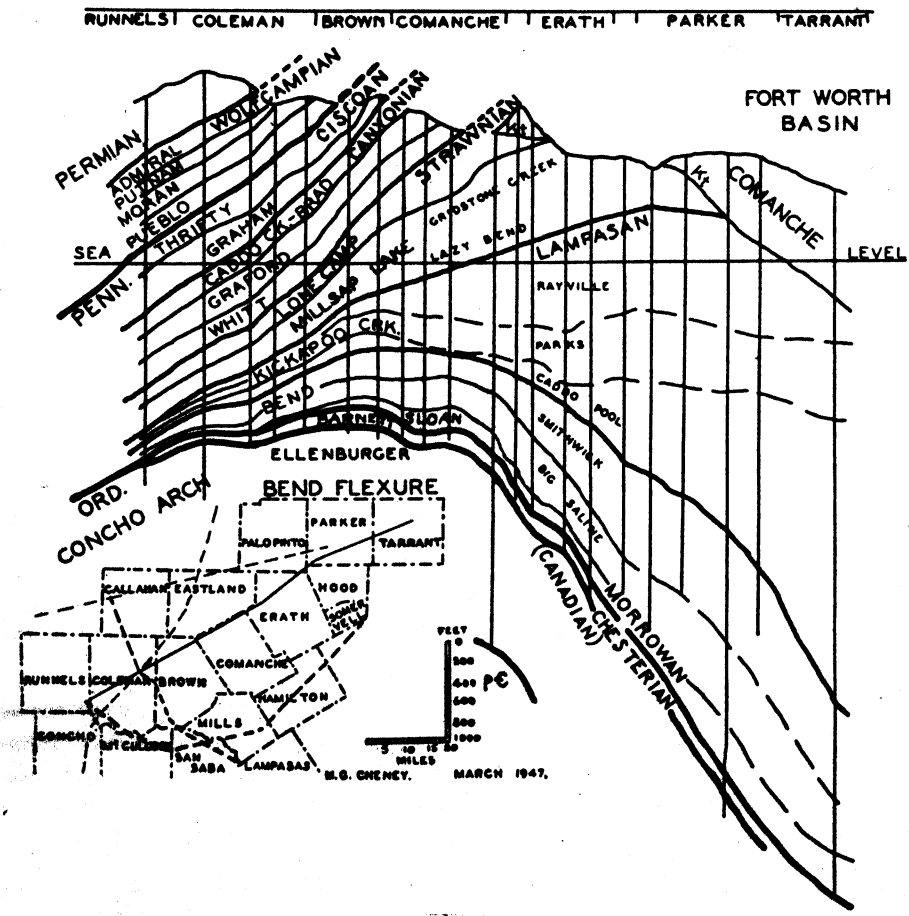

FIG. 2.—Profile from Concho arch to Fort Worth basin, showing time-stratigraphic divisions of system and series rank; also lithologic groups and certain formational units from Middle Strawn to the Upper Mississippian Barnett shale and Lower Ordovician Ellenburger limestone (see *Bull. Amer. Assoc. Pet. Geol.*, Vol. XXIX [1945], pp. 154–55, for stratigraphic details).

named 100–125-foot section of shale, dolomitic limestones, and sandstone; the Kickapoo Creek limestone; and the Dickerson shale—a combined thickness of about 200 feet. Prior publications have given brief descriptions and partially recorded the faunas of these beds.[16]

The Rayville formation thins rapidly westward from its type area, being less than 300 feet thick in Brown and Coleman counties. This formation is absent over the more positive areas, such as the Richland Springs axis, the southeastern part of the Concho arch, and the Electra and Muenster arches.

The general character and thickness of the Lampasas series of the Colorado River district is shown in Figure 3, which also illustrates its unconformable relationships with both the overlying Strawn series and the underlying Morrowan. The late Dr. F. B. Plummer assigned the name "Sloan"[17] to the outcropping Morrowan rocks, and this name should now replace the subsurface name "Comyn."

The name "Mable Falls limestone" apparently is too fixed and useful as a broad lithologic term for the Sloan and Big Saline limestone sequence to be restricted to limestones either above or below the Morrow-Lampasas series boundary. This is also true for the name "Bend group." When applied to the outcrop section, the Bend group generally includes the Sloan, Big Saline, and Smithwick. Traced underground, each of these three divisions shows sufficient rock diversity and thickness to qualify as a mappable unit. No less than formational rank is thought to be appropriate for each.

LOWER PENNSYLVANIAN

Rocks of Springer age have not been recognized at the surface in north-central Texas; those correlated with the Morrow crop out only in the eastern two-thirds of the Llano Uplift region. Recent studies of fusulinids[18] and corals[19] indicate that the upper part of the type Marble Falls and Bend divisions are Lampasan and the basal part Morrowan.

MIDDLE PENNSYLVANIAN

Separation of Sloan and Big Saline beds is difficult in areas of outcrop, because the boundary commonly falls within a limestone sequence. However, to the north and northwest, wells in many areas have disclosed the presence of chert conglomerates or coarse sandstone lenses beneath *Fusulinella*-bearing beds of the Big Saline and above oölitic or dense Pennsylvanian limestones, in which *Fusulinella* are not found. This conglomerate and sandstone zone expands northward and is correlated with the Bend conglomerate pay zone of Montague County and the Bostwick conglomerate of the Ardmore Basin in southern Oklahoma. A considerable thickness of Dornick Hills beds below the Bostwick is generally accepted as belonging to the Morrow series.

[16] Frederick B. Plummer and Raymond C. Moore, "Stratigraphy of the Pennsylvanian Formations of North-central Texas," *Univ. Tex. Bull. 2132* (1921), pp. 70–74; F. B. Plummer and J. Hornberger, Jr., "Geology of Palo Pinto County," *Univ. Tex. Bull. 3534* (1935), pp. 15–16, 19–20; Carl O. Dunbar, Lloyd G. Henbest, and J. Marvin Weller, "Pennsylvanian Fusulinidae of Illinois," *Ill. Geol. Surv. Bull. 67* (1942), pp. 113–15.

[17] "Stratigraphy of Lower Pennsylvanian Coral-bearing Strata of Texas," *Univ. Tex. Pub. 4401* (1944), pp. 66–68.

[18] M. L. Thompson, "Stratigraphy and Fusulinids of Pre-Desmoinesian Pennsylvanian Rocks, Llano Uplift, Texas," *Jour. Paleon.*, Vol. XXI (1947), pp. 147–64.

[19] Raymond C. Moore and Russell M. Jeffords, "Description of Lower Pennsylvanian Corals from Texas and Adjacent States," *Univ. Tex. Pub. 4401* (1944), pp. 77–81.

Black shales and thin limestones of Upper Big Saline age, below the olive-brown and upper black shales of the Smithwick, occur in isolated areas on the eastern and northeastern flank of the Llano Uplift and are typically developed near the village of Bend. The Caddo limestone facies of the upper "subsurface Bend" or Kickapoo Creek group extends westward from Brown, Eastland, Stephens, Young, and Jack counties. Well data show that these limestone beds grade to shale eastward. The eastern margin of this limestone–dark shale transition belt can be traced as far south as northeastern McCulloch County in the subsurface. This shale is absent, presumably by erosion, over the Richland Springs axis but reappears as a shale and sandstone sequence farther east, as shown in wells 10–17 of Figure 3.

Figure 3 also illustrates (1) the westward loss of rocks of Morrow and Springer ages, as proved by the contact of *Fusulinella*-bearing beds with Mississippian and Ordovician in well 4 and westward; (2) the variable lithologic character of the Big Saline and Smithwick formations; (3) pronounced changes in thickness of Upper Lampasas rocks as a result of structural changes near the close of Lampasan time; (4) evidence that the pronounced Llano fault system, which extends northeastward beneath the Strawn series, was developed near the close of Kickapoo Creek time and prior to deposition of the thick sandstones found in the upper part of wells 9–11 of Figure 3. As traced northward and eastward in wells, these sandstones belong to the Millsap Lake group (restricted).

The progressive overlap observed in outcrops in San Saba County extends southwestward across McCulloch to eastern Menard and Kimble counties. Strawn and Lower and Middle Canyon beds rest on Big Saline and Ellenburger limestones over the Concho arch around the western border of the present Llano Uplift.

The outcropping and subsurface Upper Lampasas beds show very different facies, according to correlations derived from Figures 3 and 4 and other regional studies.

The clastics deposited east of the Richland Springs axis probably came from the southwestern extension of the folded and uplifted Ouachita trough, with its thick sandstone, shale, and chert sequence, whereas the calcareous sediments in the V-shaped area northwest from the Richland Springs axis and northeast from the Concho arch may represent deposition on a bank or shallow platform near the mouth of northward drainage from Ellenburger limestone and dolomite surfaces which were undergoing gradual elevation along these two broad structural axes. Major faults ing and uplift of these axes appear to have occurred near the close of Lampasa-time.

Subsurface studies show that the Richland Springs axis is part of an extensive anticline which developed mainly during pre-Strawn time across what is now Mason, McCulloch, and San Saba counties to central Brown County. Farther northward this anticlinal axis became a flexure. This conclusion is based on loss of Upper Lampasas and Lower Strawn in the southern part of this major structural trend and upon no such loss of section in the northern part. Distinctive names may prove useful for different parts of this extensive feature, as follows: (1) "Richland Springs axis" for that part which may be seen at the surface in western San Saba County and for 50 miles to the southwest; (2) "Bend

arch" for the subsurface feature which developed as an arch for a distance of about 30 miles from the Richland Springs to the Brownwood district; and (3) "Bend flexure" for 120 miles northward from the Brownwood district, the present arched appearance of the Bend flexure having resulted from westward tilting during post-Pennsylvanian time.

As previously reported,[20] fusulinids from the Upper Caddo or "subsurface Bend" limestone in Coleman County indicate time equivalence with the Kickapoo Falls and Dennis Bridge limestones. As traced northward in wells, the Big Valley member (No. 12) of the Colorado outcrop section appears correlative with the Kickapoo Creek group. The area in which Upper Lampasas muds or shales were eroded and replaced by Strawn sands extends over several hundred square miles west of the major rifting and in trend with the pronounced Richland Springs structural axis. These data support the conclusion that considerable uplift and erosion must have taken place in north-central Texas near the close of the Lampasas epoch. Similar structural developments evidently occurred in eastern Oklahoma, where the Boggy, but not higher beds, was affected by the closing phases of the Ouachita orogeny. The Boggy beds are involved in pronounced northeast-trending faults with displacements up to 600 feet, as at Muskogee,[21] and sharp folds with dips of more than 20 degrees in the McAlester Basin. Choice of a basal boundary for the Des Moines series should depend in part on the structural history of the Mid-Continent region. The analysis by Shepard W. Lowman[22] deserves repeating:

Basal Marmaton
Upper Cherokee: Calvin Sandstone—Thurman sandstone
Middle Cherokee: Boggy formation—Lower Savanna sandstone
Lower Cherokee: McAlester shale—Atoka formation

Major changes in the regional structure of the mid-continent area seem to have been initiated between the middle and upper Cherokee as evidenced by the following data:

1, There is a broad regional unconformity at the base of the upper Cherokee. 2, Comparatively coarse sediments in the Chelsea-Thurman sandstone grade rapidly into finer sediments in the underground to the west indicating that the present line along which these sediments crop out is near their original shore line. 3, The area of maximum sedimentation in the upper Cherokee shifts many miles northwest of the axis of the middle and lower Cherokee basins. 4, The line of strike changes with the advent of the Thurman. 5, The folds of the "Brown lime" (note: upper Savanna) in the Seminole area show that the folding during middle Cherokee time was in phase with lower Cherokee. On the other hand the structural expression of the "Senora lime" (upper Cherokee) is disconformable with the middle and lower Cherokee. 6, The southwestern extension of the Ozark Mountains into Oklahoma appears to have been elevated in upper Cherokee time. 7, Continued uplift of many and possibly all major local uplifts seem to have persisted with diminishing intensity through upper Cherokee time and on through Marmaton, Kansas City, Lansing, and Douglas time and many of the structures are well expressed in surface strata as young as Permian. However, the regional structures like the Henryetta Flexure and the Central Oklahoma arch appear to have ceased moving with the end of the middle Cherokee time and all local folds are buried by upper Cherokee sediments. . . . .

The Hunton arch (proper) with attendant features such as the Henryetta Flexure, the Seminole Uplift, the Paul's Valley Arch, and the Central Oklahoma Arch and almost all important local "structures" had received their

[20] P. 162 of ftn. 2.

[21] Charles W. Wilson, Jr., and Norman D. Newell, "Geology of the Muskogee-Porum District, Oklahoma," *Okla. Geol. Surv. Bull. 57* (1937), p. 80.

[22] "Cherokee Structural History in Oklahoma," *Tulsa Geol. Soc. Digest* (1933), pp. 31 and 34.

essential form by the end of the Middle Cherokee time and were only modified by later events.

The Upper Cherokee and Marmaton are thought to be equivalent to the 2,000-foot section extending from the top of the Dennis Bridge limestone to the top of the East Mountain shale of the Brazos River district in north-central Texas. The Brannon Bridge, Santo, Goen, and Capps limestone members of this section show greatest lateral persistency. The genus *Wedekindellina* is rare or absent in the Goen limestone and higher beds, according to several fusulinid specialists. Previous extensive studies of the Strawn have solved many of the earlier problems of correlation, although doubtless much more remains to be learned about faunal zones, lithologic changes, and how much and what part of the loss of section in the more positive areas was caused (1) by lateral thinning when deposition was going on or (2) by truncation following uplift.

UPPER PENNSYLVANIAN

Most limestone members of the Canyon and Cisco series show considerable persistency in outcrop and subsurface. Where these Upper Pennsylvanian rocks crop out, the Comanche overlap narrows sufficiently to permit nearly continuous mapping of individual members across north-central Texas. Tracing by Plummer and Moore of the Home Creek limestone northward from its type area is believed to be more correct than the more recent interpretation by W. Lee and C. O. Nickell.[23] Other problems of the Upper Pennsylvanian are discussed below.

[23] "Stratigraphic and Paleontologic Studies of the Pennsylvanian and Permian Rocks in North-central Texas," *Univ. Tex. Pub. 3801* (1938), pp. 108–17, Pl. VIII.

SUBSURFACE STRATIGRAPHY

Classification and correlation studies of the subsurface Pennsylvanian rocks in north-central Texas must deal with a composite section about 10,000 feet thick,[24] as against 5,000 feet for the exposed sections and about 1,000 feet over the Concho, Electra, and Muenster arches. Pennsylvanian rocks outcrop or are reached at a depth of less than 1 mile throughout an area of about 70,000 square miles. Subsurface rock samples are available from many of the tens of thousands of wells drilled for petroleum in this region.

LOWER PENNSYLVANIAN

Rocks of Springer age have been identified in well cuttings by H. D. Miser and E. H. Sellards[25] in the extreme eastern portion of this region, evidently proving that the early Pennsylvanian Ouachita Basin of southeastern Oklahoma extended southwestward into Texas. Quite possibly this deeply subsiding trough continued around the Concho arch to connect with the Tesnus trough of the Marathon region. The Springer series is thought to be of the same age as the Pocahontas series (Lower Pottsville) of the northern Appalachian standard section.

The Sloan formation of Morrow age is present in the basin areas and over the Bend flexure (or arch) but is absent over the Concho, Electra, and Muenster arches.

[24] Pp. 154–55 and Fig. 2 of ftn. 1.

[25] "Pre-Cretaceous Rocks Found in Wells in Gulf Coastal Plain South of Ouachita Mountains," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XV (1931), pp. 801–18; E. H. Sellards, "Rocks Underlying Cretaceous in Balcones Fault Zone of Central Texas," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XV (1931), pp. 819–27.

As previously stated, the original usage of the name "Marble Falls limestone" as defined by Hill may well be retained as a broad lithologic term to be used where differentiation into Sloan and Big Saline is not established. The Morrow series is closely equivalent to the New River series (Middle Pottsville) of the northern Appalachian standard section.

MIDDLE PENNSYLVANIAN

The two most significant and pronounced changes in the Pennsylvanian fauna in the Mid-Continent occur at the Morrow-Lampasas and the Des Moines–Missouri series boundaries. The strata between these two outstanding boundaries attain a thickness of at least 15,000 feet in the McAlester and Atoka basins in southeastern Oklahoma. It is now generally accepted that this thick section should be divided into two series rather than be included in the Des Moines series, as was the custom until the late 1930's. The name "Winslow" was used to some extent for the Atoka plus a variable amount of overlying section but was formally discarded by the United States Geological Survey because of varied usage. Many geologists in Oklahoma have expressed objection to the revival of "Winslow" as the name for a series between the Morrow and the Des Moines series. As early as 1935 some unpublished reports began dividing this thick section into Atoka and Des Moines series because of faunal, lithologic, and structural changes at the top of the Atoka. Apparently, the first proposal of two series to appear in formal literature was in 1940 when the name "Lampasas" was introduced for a post-Morrow–pre-Des Moines series.[26]

[26] Pp. 81–87, Fig. 1, of ftn. 5.

Doubtless, the identification[27] of plants of the *Neuropteris rarinervis* zone of Clarion age in the Hartshorne-Boggy formations played some part in the inclusion of these post-Atoka beds in the Des Moines series, because most of the Clarion rocks were then commonly included in the Allegheny division in Pennsylvania and Ohio. The recently expressed preference by the state geologist of Pennsylvania and paleobotanical specialists for an Allegheny-Kanawha (Upper Pottsville) boundary at or very near the top of the Clarion beds supports the selection of a comparable boundary higher than the top of Atoka in the Mid-Continent region.

The Lampasas series, as defined for north-central Texas, was thought properly to include all post-Morrow rocks commonly assigned to the Bend group of the subsurface. Geological literature is replete with papers dealing with the geology of the pre-Strawn gray and black limestone and black shale sequence of the subsurface.[28] It was recognized that the 600-foot upper limestone section of this 1,200-foot sequence changes eastward

[27] T. A. Hendricks and C. B. Read, "Correlations of Pennsylvanian Strata in Arkansas and Oklahoma Coal Fields," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XVIII (1934), p. 1046 n.

[28] Wallace E. Pratt, "Geologic Structures and Producing Areas in North Texas Petroleum Fields," *Bull. Amer. Assoc. Pet. Geol.*, Vol. III (1919), pp. 44–70; W. G. Matteson, "A Review of the Development in the New Central Texas Oil Fields during 1918," *Bull. Amer. Assoc. Pet. Geol.*, Vol. III (1919), pp. 163–211; Plummer and Moore, p. 42, Pl. 5, of ftn. 16; E. K. Esgen, "Relation of Accumulation of Petroleum to Structure in Stephens County, Texas," *Structure of Typical American Oil Fields*, Vol. II (Amer. Assoc. Pet. Geol., 1929), pp. 470–79; Frank E. Kendrick and Philip G. Russell, "Natural Gas in Bend Arch District, Texas," *Geology of Natural Gas* (Amer. Assoc. Pet. Geol., 1935), pp. 609–49; Fort Worth Geological Society, H. H. Bradfield, *et al.*, "Geologic Cross-Section of North-central Texas" (Dec. 21, 1940).

into black and dark shales, whereas westward it is much reduced in thickness by thinning and truncation toward the Concho arch. Likewise, it has been widely held that the structural features of these subsurface Bend rocks differ notably from those observed in overlying beds. Thus many writers believed that important differences of lithology and structure clearly differentiated the Caddo limestone and black-shale section from the overlying sandstones and shales with some interbedded limestones, which typify the Strawn division.

In recent years fusulinid specialists have noted the presence of *Wedekindellina* and early types of *Fusulina* s.s. in the Caddo limestone and have recommended extending the Strawn series down to the base of the Caddo Pool formation. In accordance with this view, Spivey and Roberts have presented a strong case for an Atoka series in both Oklahoma and north-central Texas.[29] There are, however, objections to the suitability of the Atoka because of structural complications in the type area. Also earliest representatives of *Wedekindellina* and *Fusulina* occur in the Owens-Frensley beds of the Ardmore Basin, probably correlative with the Red Oak zone, 500 feet below the top of the Atoka.[30] Upper Kanawha and Tradewater beds of the Appalachian and Illinois basins are doubtless of post-Atoka age.

Perhaps a portion of the outcropping Upper Smithwick shale near Smithwick may be equivalent to the siliceous Caddo Pool formation of the subsurface. The Smithwick increases in thickness eastward and becomes much more sandy near its original type locality than in areas farther west and northwest.

The predominantly limestone development of the Kickapoo Creek group in the Bend flexure region seems to be an example of biostrome development.[31] These rocks have yielded more than 200,000,000 barrels of oil in Stephens, Young, Shackelford, Eastland, and adjoining counties.

The name "Bend" was used in formal literature from the 1890's to about 1918 to designate outcropping rocks ranging from the base of the Barnett to the top of the Smithwick. Since 1918 it has been applied the Marble Falls and Smithwick of the outcrop, and in the subsurface to the Mable Falls, Smithwick, and younger strata now assigned to the Kickapoo Creek group. Recently, the use of "Bend" as a series name restricted to the Big Saline and Smithwick of the outcrop section has been advocated.[32] Thus the names "Winslow," "Atoka," "Lampasas," and "Bend" have been advocated for a post-Morrow–pre-Des Moines series in the Mid-Continent region. The best solution of these conflicting views and usages is not readily apparent. However, if the Clarion rocks of the type Pennsylvanian section are assigned to the Kanawha series then the "Lampasas series," as used herein, is to be preferred on the basis of time equivalency as judged by present paleontologic evidence.

The type Kanawha division of West Virginia has, for the last forty years, included in its upper part the Kanawha black flint and Roaring Creek sandstone, which H. R. Wanless[33] tentatively

[29] Pp. 185–86 of ftn. 6.

[30] H. H. Bradfield, personal communication, January 15, 1947.

[31] E. R. Cumings, "Reef or Bioherms?" *Bull. Geol. Soc. Amer.*, Vol. XLIII (1922), pp. 331–52.

[32] P. 73 of ftn. 17.

[33] "Pennsylvanian Correlations in the Eastern Interior and Appalachian Coal Fields," *Geol. Soc. Amer. Spec. Paper 17* (1939), p. 50.

correlated with the Zaleski flint and Clarion sandstone, respectively, above the Putnam Hill limestone of eastern Ohio and western Pennsylvania. M. L. Thompson[34] reported *Fusulina leei*[35] in the Putnam Hill limestone and the transitional species, *Fusulinella* or *Fusulina carmani*, in the Vanport limestone which Ashley[36] places in the Pottsville. These or similar primitive *Fusulina* species occur in upper Tradewater and Boggy beds.

M. T. Sturgeon reports that the striated forms of *Mesolobus* extend up to, or nearly to, the Vanport limestone, whereas the smooth types appear abundantly in the Hamden limestone of the lower Kittanning section of eastern Ohio.[37] A. T. Cross has found that "megaspore assemblages" indicate that "major floristic changes" occur between the Coalburg and the Buffalo Creek coals of West Virginia, somewhat below the top of the Kanawha division.[38] According to Ashley's correlations,[39] this floral break is somewhat above the Brookville and below the Lower Kittanning coal of Beaver Valley, Pennsylvania.

Correlations of these rocks with the Illinois and Mid-Continent section have been made by plants, fusulinids, brachiopods, and other fossils. As previously stated, Read concluded that Clarion plants are present in the Hartshorne, McAlester, and Savanna and as high as studied in the Boggy of the McAlester Basin. No paleobotanical data seem to have been gathered for this part of the Pennsylvanian system in north-central Texas. Dunbar, Henbest, and Weller have reviewed correlations between the northern Appalachian region and Illinois based on fusulinids and other faunas.[40] They report that the advanced type of *Fusulinella*—*F. iowensis*—"appears to have a rather limited stratigraphic range," being "known only in Seville limestone in Illinois, at a single horizon in Iowa (90 feet below the Whitebreast coal) that is probably equivalent to the Seville, and in the Upper and Lower Mercer limestones of Ohio." The Mercer limestones have invariably been classified as Upper Pottsville in the Appalachian region. Species presumably included in the subzone *Fusulina leei* are recorded from the Clarion, Upper Tradewater, Savanna, Lower Boggy, Upper Dornick Hills, and Kickapoo Falls of Ohio, Illinois, Oklahoma, and north-central Texas, respectively. Only one species, *Wedekindellina euthysepta*, is shown in this report to cross the Tradewater-Carbondale boundary in Illinois.

It is doubtless significant that approximately equivalent boundaries near the middle of the Pennsylvanian system were chosen independently by early workers in the northern Appalachian, Illinois, Ardmore, and north-central Texas (subsurface) regions. Past differences in classification of this part of the section in the northern Mid-Continent may be attributed, perhaps, to the view that the Des Moines series by definition must include all pre-Missourian Pennsylvanian found in the Des Moines River region. However, the oldest Pennsylvanian rocks of the Des Moines River section contain fusulinids of the *F. iowensis* and *F. leei*

[34] "Pennsylvanian Fusulinids from Ohio," *Jour. Paleon.*, Vol. X (1936), pp. 677–79.

[35] Dunbar, Henbest, and Weller, p. 30, of ftn. 16.

[36] George H. Ashley, "The Pittsburgh-Pottsville Boundary," *Jour. Geol.*, Vol. XXIX (1945), pp. 125–69.

[37] "Allegheny Fossil Invertebrates from Eastern Ohio–Nautiloidea," *Jour. Paleon.*, Vol. XX (1946), pp. 11–12.

[38] "Spore Floras of the Pennsylvanian of West Virginia," *Bull. Geol. Soc. Amer.*, Vol. LVII (1946), p. 1265.

[39] Fig. 5, p. 381, of ftn. 11.

[40] Pp. 28–34 of ftn. 16.

subzones and other fauna characteristic of the Upper Pottsville. Therefore, it may prove more useful for purposes of interregional correlation to recognize the lower boundary of the Des Moines series in the northern Mid-Continent region as occurring between the zones of early and late types of *Fusulina* instead of between the restricted and advanced types of *Fusulinella*.

Drilling and geophysical work have revealed that Bend and older rocks in the subsurface are affected by an extensive system of normal faults or steep folds with northeast trends, like those observed at the surface in the Llano Uplift region. As previously mentioned, these structures resemble the northeast-trending fault system of the Muskogee, Oklahoma, district, where normal faults with as much as 600 feet of displacement cut the Bluejacket sandstone of the lower Boggy.[41] Upper Cherokee and Millsap Lake beds exhibit much less structural disturbance than do older rocks, and their structural features commonly trend northwest instead of northeast.

UPPER PENNSYLVANIAN

Correlations of Canyon and Cisco rocks throughout north-central Texas are well established, with the possible exception of the Home Creek limestone as noted above. Lowering of the basal Canyon boundary has been proposed in recent years to accomplish closer agreement with the Missouri–Des Moines boundary of the northern Mid-Continent region.

W. F. Cummins[42] quite logically drew a lithologic group boundary at the base of the thick Palo Pinto limestone in his early study of the Brazos River Valley section. Subsequently it has been found that this thick limestone grades into shale northward in the subsurface and southward in the outcrops of the Colorado River section. The Palo Pinto is not equivalent to the Capps limestone in the latter area, as formerly believed, but, if present, it occurs near the middle of the 300-foot Brownwood shale section. Representatives of the genera *Fusulina* and *Mesolobus* are present in the Capps limestone (but not much, if any, higher) in the Brownwood district. These Des Moines fossils occur only as high as the East Mountain shale of the Mineral Wells district. Obviously, a series boundary based on faunal changes and prominent disconformity can be far more useful for careful subsurface geological work than was the former lithologic boundary based on a limestone of rather local extent. Thus in the Mineral Wells district the base of the Lake Pinto conglomerate should serve as a Strawn-Canyon series boundary more satisfactorily than the base of the Palo Pinto limestone, some 500 feet higher in the section.

The long-established boundary between Canyon and Cisco beds apparently requires no revision in order to qualify these divisions as series rather than as groups. The Kisinger channel,[43] which cuts through the Home Creek and down to the Ranger limestone in southeastern Young County, is one of many indications of a pronounced hiatus subsequent to Home Creek deposition. Cavey red shales and numerous areas of truncation of the Home Creek beds occur widely in north-central Texas. Canyon and older beds below this hiatus tend to thicken eastward, whereas Lower Cisco beds thicken westward and northwest-

[41] Wilson and Newell, pp. 74–84 of ftn. 21.

[42] "Report on the Geology of Northwestern Texas," *Tex. Geol. Surv., 2d Ann. Rept.* (1890), Pls. XVI, XVIII.

[43] Lee and Nickell, pp. 12–16, Pl. IV, Fig. 2, of ftn. 23.

ward from its area of outcrop in the Colorado River Valley.

Like the subsurface Bend group, the Canyon series is characterized by extensive calcareous deposits, which are difficult to correlate in some localities because of rather abrupt transitions. Further studies are needed to understand more fully the localization of these limestone deposits. The 200-foot limestone of the Palo Pinto formation in the subsurface of Stephens County grades into shales in northern Young and southern Eastland County, but the somewhat younger 125-foot Winchell limestone of Brown and Eastland counties becomes very thin northward in Young County. The Ranger limestone is relatively thick near its outcrop from Wise to McCulloch counties, but it thins westward in many areas. The total thickness of the Canyon series is not particularly affected by these lithologic changes. More or less extensive bioherms or reefs are quite common and in some areas extend upward throughout much, or even all, of the Canyon section. Some of these thick, calcareous bodies are dolomitized, with sulphur water, oil, or gas filling the porous zones. Many calcareous deposits of the Canyon series appear to be so local as to qualify as bioherms, others sufficiently extensive to resemble the biostromes of the Upper Lampasas series. Canyon limestone rests on Upper Strawn limestones and is followed by Cisco limestones on the Petrolia Uplift of Clay County and westward along the Electra arch. In earlier correlations[44] the lower portion of Pennsylvanian reef deposits was misidentified as Ellenburger, which it overlies in the Petrolia and Electra districts. Farther southward the limestone masses do not appear to be closely related to local structural features. They were evidently developed in shallow seas on banks or slightly raised platforms considerably removed from sources of clastic material. The seas of Canyon times covered all the more positive areas of north-central Texas, including the Muenster, Electra, and Concho arches.

Downward revision of the upper boundary of the Cisco series and Pennsylvanian system appears desirable, in order to reach agreement with the Permian boundary established in West Texas and elsewhere.[45] Permian schwagerinids occur down to, or perhaps even below, the Waldrip No. 2 limestone. Pronounced channels between several marine limestone members occur in this part of the section, indicating that elsewhere marked crustal disturbance was in progress, presumably the late Marathon thrust movements and the concluding phases of the Arbuckle-Wichita orogeny, which had been active during much of Cisco time.

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[44] M. G. Cheney, "Stratigraphic and Structural Studies in Northcentral Texas," *Univ. Tex. Bull. 2913* (1929), Pl. V.

[45] C. W. Tomlinson *et al.*, "Classification of Permian Rocks," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XXIV (1940), pp. 337–42, 352–54.

DISTRIBUTION AND CORRELATION OF PENNSYLVANIAN ROCKS IN LATE PALEOZOIC SEDIMENTARY BASINS OF NORTHERN NEW MEXICO[1]

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ABSTRACT

The most useful classification of the Pennsylvania series in northern New Mexico is by the lithologic character of the rocks, supplemented by paleontologic data where available. Few of the boundaries are sharp. The strata represent a single major sedimentary cycle, comprising, in ascending order, a suite of transgressive sediments, a widely distributed marine phase, and an unevenly distributed regressive phase, followed by continental sediments. For the marine part of the sequence the term "Magdalena group" is used, including the Sandia formation and the Madera limestone, the latter containing several named members. The positions of major positive areas and of depositional basins, total thicknesses, and general character of the sediments are indicated on a map of northern New Mexico, and detailed local sections are shown graphically in several charts.

INTRODUCTION

Field parties of the U.S. Geological Survey have been engaged in mapping parts of northern New Mexico for several years, much of the work being in areas of exposure of Pennsylvanian strata. The combined projects have covered approximately 16,000 square miles, and investigations are still in progress. This report is intended to summarize progress to date but is not a valedictory on the Pennsylvanian series in northern New Mexico.

The following individuals have participated with the authors in the investigations in New Mexico as geologists of the Geological Survey: Ernest Dobrovolny, V. C. Kelley, A. J. MacAlpin, S. A. Northrop, C. P. Rogers, Jr., H. H. Sullwold, Jr., C. H. Summerson, and R. H. Wilpolt. In addition, R. L. Bates and Georges Vorbe, of the New Mexico Bureau of Mines and Mineral Resources, have actively co-operated in several of the investigations and have offered much valuable counsel.

[1] Published by permission of the director, U.S. Geological Survey.

THE PENNSYLVANIAN SERIES IN NORTHERN NEW MEXICO

The Pennsylvanian series in New Mexico is composed of a complex sequence of marine and continental strata that has a lobate geographic distribution as a result of the presence of several Pennsylvanian positive elements. The rocks are exposed in belts of outcrop in and adjacent to the principal mountain ranges of the state and are known to be present at depth under several of the intermountain valleys, as well as under portions of the Colorado Plateau and High Plains provinces.

Regional analysis of the Pennsylvanian series in New Mexico has required much preliminary planning. The results of the work directly reflect these plans, and we think it essential for a proper understanding of the ensuing discussion to disgress briefly from our main objective to present our approach to the problem.

METHODS OF STUDY

Proper planning of a regional study consists in defining the objectives, determining the critical and the secondary

areas, and reaching decisions on methods and classifications to be used. It was clear that the principal objective for the purposes of this report was a study of the regional distribution and stratigraphy of the Pennsylvanian series. Critical areas were the belts of outcrop in and adjacent to the mountain uplifts in northern New Mexico; secondary, because of lack of information at the present time, were vast areas in which the Pennsylvanian strata are covered by representatives of later series and systems. Since much of New Mexico is structurally complex, it was clear that stratigraphy and structural geology must be considered together in mapping. This has provided much opportunity for examination of lateral stratigraphic variations that are often neglected in usual stratigraphic procedures.

The basis for a stratigraphic classification is either paleontologic or lithologic. Neither is completely independent, but clearly a decision must be reached early in such projects as the one presented here regarding which should be emphasized and what criteria should be employed to determine boundaries of units.

Stratigraphic classification based primarily on paleontologic criteria requires either much prior information on faunal sequences in the area under study and in other important areas of similar strata, or there must be time available for such study concurrent with the other phases of the stratigraphic work. Prior information on faunal sequences in the Pennsylvanian of New Mexico was not available; there was neither the time nor the facilities to undertake such work in conjunction with the assignment; and, finally, we were not impressed by the adequacy of such information in other areas with which comparison might be necessary. At the beginning of the work we were—and we still are—doubtful of the practical merits of mapping, or even the possibility of accurately identifying, the limits of faunal zones in series of the general type represented by the Pennsylvanian.

Development of a lithologic classification for the Pennsylvanian series in northern New Mexico is a complex problem. It requires decision as to procedure in defining major and minor divisions that directs attention to the evaluation of the importance of lithologic changes and disconformable relations. Even though lithologic character is the fundamental basis selected, it cannot be divorced from paleontology in problems of correlation, and therefore concurrent studies of the fossil content of the rocks must be undertaken.

Physical and theoretical evidence of disconformities and unconformities has been a major criterion whereby the Pennsylvanian series in other parts of the United States has been divided into its several units.[2] The use of such bases harks back to the opinion that diastrophic events of regional importance provide an adequate and very natural foundation for stratigraphic division. With such opinion we are in full accord, but we suspect that overuse of this method in the Pennsylvanian series has resulted in the development of a pseudo-uniform classification that does not agree with the observed variations throughout the North American continent.

It appears to us probable that many disconformities are local; that the physical evidence of disconformity, or of unconformity, in many instances is no measure of the time involved and affords no criterion whereby a major

[2] R. C. Moore and Others, "Correlations of Pennsylvanian Formations of North America," *Bull. Geol. Soc. Amer.*, Vol. LV (1944), pp. 657–705.

hiatus may be distinguished from a very minor, unimportant local break; and that paleontologists have often failed to discriminate between faunal change and faunal break. In consequence, we suspect that some disconformities and hiatuses have not been traced but have been postulated beyond the limits of their actual development.

We believe that sedimentation in many geosynclinal and foreland areas during Pennsylvanian time was nearly or quite continuous during periods of mountain-building marginal to such features. Observation in the field also leads to the conclusion that disconformities and unconformities in Pennsylvanian rocks in northern New Mexico most commonly exist on the margins of geanticlines, decrease in magnitude basinward, and, finally, are lost in continuous sequences of conformable strata in the basin areas.

Clearly, the measure of the magnitude of apparent hiatuses inferred from physical evidence is the degree of discontinuity of the biotas. Absence of biotas and not changes in assemblages due to processes of evolution, migration, or facies shift is evidence of a hiatus. To illustrate, the appearance of *Callipteris conferta* and some of its associates in strata of Permian(?) age is not a criterion of a hiatus between the lowest occurrence of such a flora and immediately subjacent rocks that contain the so-called *Odontopteris* flora. The organic succession is normal and characteristic of many sequences of late Paleozoic rocks. In alternations of marine and continental rocks the occurrence of primitive species of *Triticites* (*T. irregularis*, etc.) above advanced species of *Fusulina*, together with physical evidence, has been locally adjudged to indicate an important and very significant stratigraphic break or disconformity. If such is the case, there should clearly be a faunal break, demonstration of which requires the discovery of one or more intermediate or "mixed" faunas in areas where the break is spanned by passage beds.

The assumption that disconformities are widespread during periods of orogeny seems unjustified unless rather complete withdrawal of seas from the continental regions is clearly shown by a great array of data. Even then it is to be expected that sedimentation may have been continuous, but with shift from marine to continental facies, unless there was further elevation sufficient to develop coastal uplands rather than lowlands. On the basis of the local geology of northern New Mexico we suggest that orogeny in geanticlinal areas was accompanied, in some cases, by major and continuous sedimentation in flanking troughs, geosynclines, and forelands.

In northern New Mexico gradual lithologic changes that reflect widespread transgressions and regressions of marine embayments and the gradual development of geanticlines and their attending geosynclines are characteristic of the boundaries of units mapped in recent work. In belts marginal to geanticlines both disconformities and unconformities are common and are often of considerable magnitude, but it has not been demonstrated that these features extend into the more distal portions of the depositional areas.

Summarizing the above discussion, we have accepted lithologic features as a principal basis in dividing the Pennsylvanian series in northern New Mexico. Paleontology is considered a guide for recognition of formations and not a method of establishing them. The procedure has been followed of limiting units

by changes in sedimentation rather than by unconformities, which, in the area in question, have not been demonstrated to be regionally continuous or to represent important hiatuses. There has resulted from this procedure, which we believe to be in accord with important principles of sedimentation, the establishment of several contacts of stratigraphic units which are not "line-thin." Instead, the boundaries in many instances have been drawn in terms of transition zones.

NATURE OF SEDIMENTATION IN THE PENNSYLVANIAN SERIES

The Pennsylvanian and Permian(?) strata in northern New Mexico constitute a major sedimentary cycle. This cycle is characterized by two phases, one marine and the other continental.

The marine phase of sedimentation consists of (1) a suite of transgressive sediments; (2) evenly and widely distributed marine sediments that were deposited during a period of maximum marine transgression; and (3) unevenly distributed and restricted alternating marine and continental sediments that represent a period of offlap or marine regression. The strata that constitute this suite were deposited during those portions of the Pennsylvanian epoch locally characterized by development of geanticlines and sedimentation in geosynclines. Marine regression, at least to a minor extent, appears to have resulted from the rapid accumulation of clastic sediments, which appears to have exceeded subsidence in the geosynclines and in the foreland areas.

The continental phase of sedimentation has not been separated into minor units. It represents the final phase of geosynclinal sedimentation for the cycle and of geanticlinal maturity and old age.

MAJOR SUBDIVISIONS OF THE PENNSYLVANIAN SERIES IN NORTHERN NEW MEXICO

For the marine portion of the sedimentary cycle outlined above it is recommended that the term "Magdalena group" be used. Such usage is in rather close agreement with earlier definitions of the term.[3] The upper limit of the Magdalena group, as the term is used here, is not a time line but represents the limit of marine sedimentation in the area. The inferred final regression of the sea in any given area in northern New Mexico is undoubtedly a boundary of considerable importance.

For continental arkoses and red beds that rest conformably, disconformably, or unconformably on the Magdalena group over large areas, several terms have been used. Throughout much of the state "Abo formation" is accepted. In the Rowe-Mora Basin, where indivisible arkose and red shale range in age from Upper Pennsylvanian to Permian, "Sangre de Cristo formation" has been used. In the northern part of the Nacimiento Mountains the term "Cutler formation," which has been widely used in the adjacent San Juan Mountains, is applied to continental arkose.

MINOR SUBDIVISIONS OF THE PENNSYLVANIAN SERIES IN NORTHERN NEW MEXICO

Although it has been considered inadvisable to attempt subdivision of the Abo, Sangre de Cristo, or Cutler formations in this area, it is possible to separate the Magdalena group into several units of formational rank and member-rank. The separation has been based on lithologic changes that reflect transgres-

[3] C. H. Gordon, "Notes on the Pennsylvanian Formations in the Rio Grande Valley," *Jour. Geol.*, Vol. XV (1907), pp. 805–16.

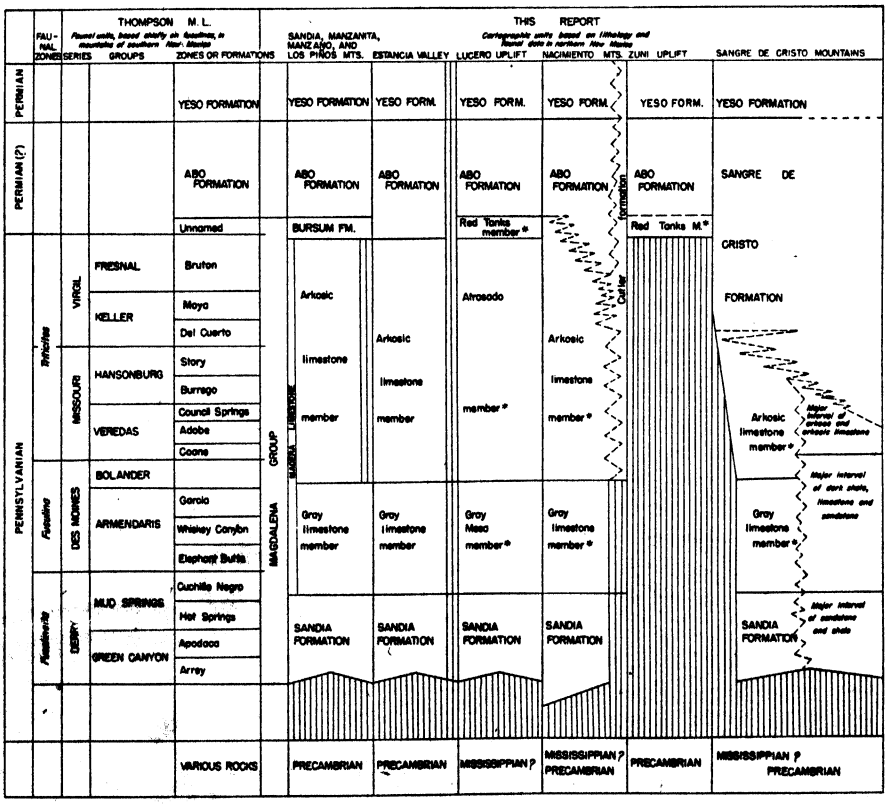
sions and regressions of the sea and periods of mountain-building.

Unconformities as unit boundaries have been taken into consideration in areas marginal to geanticlines, but such features do not provide sound criteria in the geosynclines. Faunal zones have been considered a local guide to formations and members.

The subdivision of the Magdalena group, according to the foregoing principles, is rather simple, for this group contains stratigraphic units thicker than in other Pennsylvanian formations and makes possible the recognition of formations and members as local units (Fig. 1).

The basal unit of the Magdalena group is the Sandia formation,[4] a dominantly clastic deposit of early Pennsylvanian age. Since it is a transgressive suite of sediments, it is evident that the unit must, in a regional sense, vary slightly in age limits. Such variation is minor, according to available data. Locally the basal sediments of the Sandia

[4] C. L. Herrick, "The Geology of the White Sands of New Mexico," *Jour. Geol.*, Vol. VIII (1900), pp. 112–18.



* *Members of Madera limestone*

CURRENT NOMENCLATURE AND CORRELATION OF PENNSYLVANIAN ROCK UNITS IN PARTS OF NEW MEXICO

FIG. 1

formation have been dated by contained fossils as of upper Morrow age.[5] The upper limits of the formation appear to lie within the *Fusulinella* zone and, in consequence, fall within what has been called the "Lampasas series" by some geologists in the mid-continent region.

The remainder of the Magdalena group, represented by dominantly marine sediments, is, in most of northern New Mexico, assigned to a single formation, the Madera limestone, which has been divided into several local members. Justification for this separation into members rather than into formations is found in the major difficulties of closely delimiting units that show gradual transition at their contacts. In the Lucero Uplift, on the margin of the Colorado plateaus, the Gray Mesa, Atrasado, and Red Tanks members of the Madera limestone have been recognized.[6] In the ranges east of the Rio Grande the Madera limestone[7] has been divided into a lower gray limestone member and an upper arkosic limestone member, and the Bursum formation has been locally mapped at the top.[8] This nomenclature has likewise been used in the Sierra Nacimiento[9] and in the southern quadrangles of the Sangre de Cristo Mountains.[10]

Farther north, in the Sangre de Cristo Mountains, equivalents of the Sandia and Madera formations show major changes in lithology and marked thickening. Since this area is now being mapped, it is not advisable to recommend a final nomenclature.

POSITION OF GEANTICLINES, GEOSYNCLINES, AND FORELANDS

It now seems desirable to consider the general distribution of major positive areas and depositional basins that characterized the region during much of Pennsylvanian time (Fig. 2). Two large geanticlines or linear positive elements manifested themselves as major features in New Mexico and greatly influenced sedimentation over large areas. The easternmost of these is the Sierra Grande arch, which enters the state from the northeast in Union County and extends southwestward for an unknown distance. The second is the Uncompahgre positive axis, which enters the state from the northwest in Rio Arriba and Taos counties, has an arcuate form, and extends at least into central Torrance County. Of less regional importance but locally influencing sedimentation to a marked degree is the Pedernal axis in western Torrance and southwestern San Miguel counties. The Peñasco or Nacimiento–San Pedro positive axis and its linear or echelon continuation, the Joyita axis, are likewise pronounced but of secondary regional importance.

[5] S. A. Northrop and G. H. Wood, "Geology of the Nacimiento and San Pedro Mountains and Adjacent Plateaus," *U.S. Dept. Int., Geol. Surv., Oil and Gas Investigation, Prelim. Map 57* (1946).

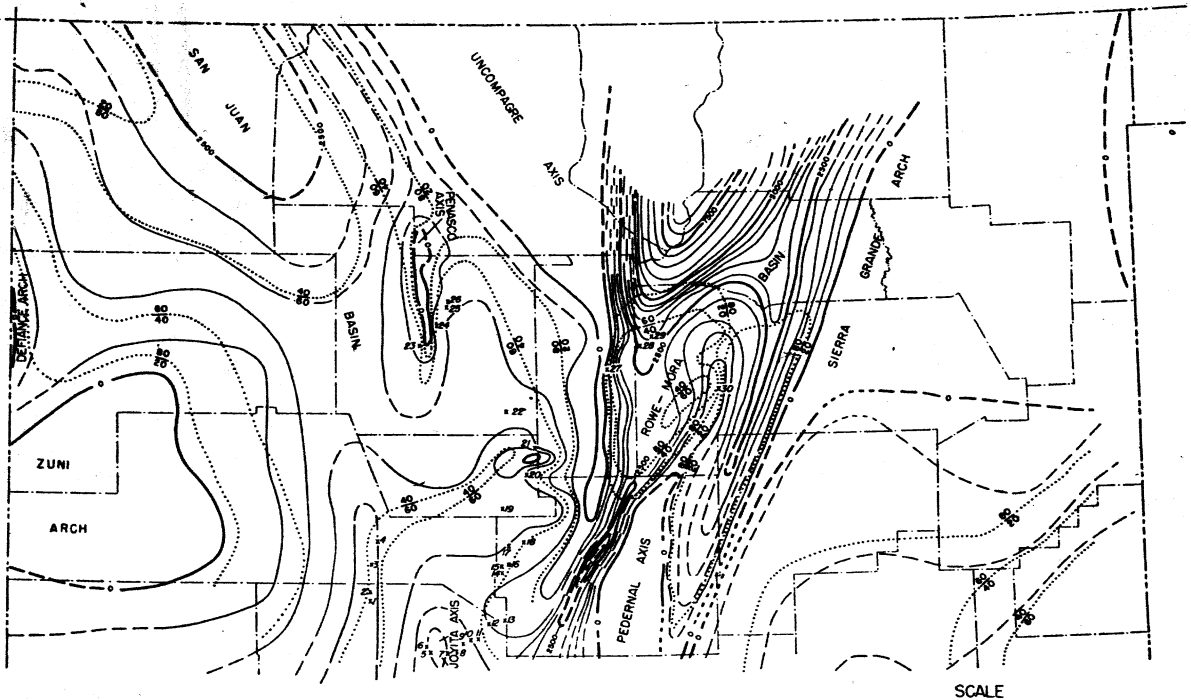
[6] V. C. Kelley and G. H. Wood, "Lucero Uplift, Valencia, Socorro, and Bernalillo Counties, New Mexico," *U.S. Dept. Int., Geol. Surv., Oil and Gas Investigation, Prelim. Map 47* (1946).

[7] C. B. Read and Others, "Geologic Map and Stratigraphic Sections of Permian and Pennsylvanian Rocks in Parts of San Miguel, Sante Fe, Sandoval, Bernalillo, Torrance, and Valencia Counties, North Central New Mexico," *U.S. Dept. Int., Geol. Surv., Oil and Gas Investigation, Prelim. Map 21* (1944).

[8] R. H. Wilpolt, A. J. MacAlpin, R. L. Bates, and Georges Vorbes, "Geologic Map and Stratigraphic Sections of Paleozoic Rocks of Joyita Hills, Los Piños Mountains, and Northern Chupadera Mesa, Valencia, Torrance, and Socorro Counties, New Mexico," *U.S. Dept. Int., Geol. Surv., Oil and Gas Investigation, Prelim. Map 61* (1946).

[9] Northrop and Wood, ftn. 5.

[10] S. A. Northrop and Others, "Geologic Maps of the Las Vegas Basin and of the Foothills of the Sangre de Cristo Mountains, San Miguel and Mora Counties, New Mexico," *U.S. Dept. Int., Geol. Surv.* (in preparation).



MAP SHOWING THICKNESS AND FACIES OF PENNSYLVANIAN ROCKS IN PARTS OF NORTHERN NEW MEXICO

SCALE

10 0 10 20 30 40 MILES

1500 — Lines of equal thickness drawn at intervals of 500 feet.

40/60 — Ratio of clastic to calcareous sediments. Upper number is clastic factor; lower is calcareous factor.

• 27 — Number and location of graphic section.

FIG. 2

The inactive arches of the Zuni and Defiance areas are near the New Mexico–Arizona state boundary. Although these features influenced distribution of Pennsylvanian sediments, they were not major contributors of clastic materials.

The major geosyncline, which has been called the "Rowe-Mora Basin" or "trough," lies between the Sierra Grande and the Uncompahgre geanticlines. Between the Uncompahgre axis and the Peñasco-Joyita axis lies a minor basin, and farther west is the ancestral or Pennsylvanian San Juan Basin, which appears to have been in part geosynclinal and in part foreland.

In addition to indicating in a general fashion the positions of geanticlines, basins, and forelands, Figure 2 shows approximate thicknesses of Pennsylvanian sediments in northern New Mexico on the basis of present available data. Since the Pennsylvanian strata of this area are dominantly of two types, clastic or calcareous, it is apparent that some expression of facies can be achieved by contouring the ratio of clastic to calcareous sediments in measured sections. By this means a quantitative expression of certain facies distribution is established.

In the San Juan Basin of western New Mexico both thicknesses and facies of the Pennsylvanian rocks are shown in a tentative fashion, with the recognition that additional data from drilling will result in revision. On the flanks of the Zuni and Defiance uplifts data are available, and in north-central New Mexico numerous belts of outcrops in and around the various mountain ranges provide much information. East of the Sangre de Cristo Mountains front, data are inadequate for other than very tentative plotting of thickness and facies lines.

Of interest in Figure 2 are gradual shifts from calcareous to dominantly clastic facies in the Rowe-Mora Basin in passing from the southern bifurcation to the broad northern portion. This major facies shift is accompanied by a marked expansion of the Lower Pennsylvanian interval. Complex facies and thickness relations of Pennsylvanian strata are, in fact, apparent throughout the entire ancestral Rocky Mountain province. To the southwest and to the west there are suggested, on the basis of widely spaced data, more gradual changes in thickness and facies. In the San Juan Basin the location of the deeper portion of the late Paleozic Basin adjacent to the Uncompahgre geanticline is evident.[11]

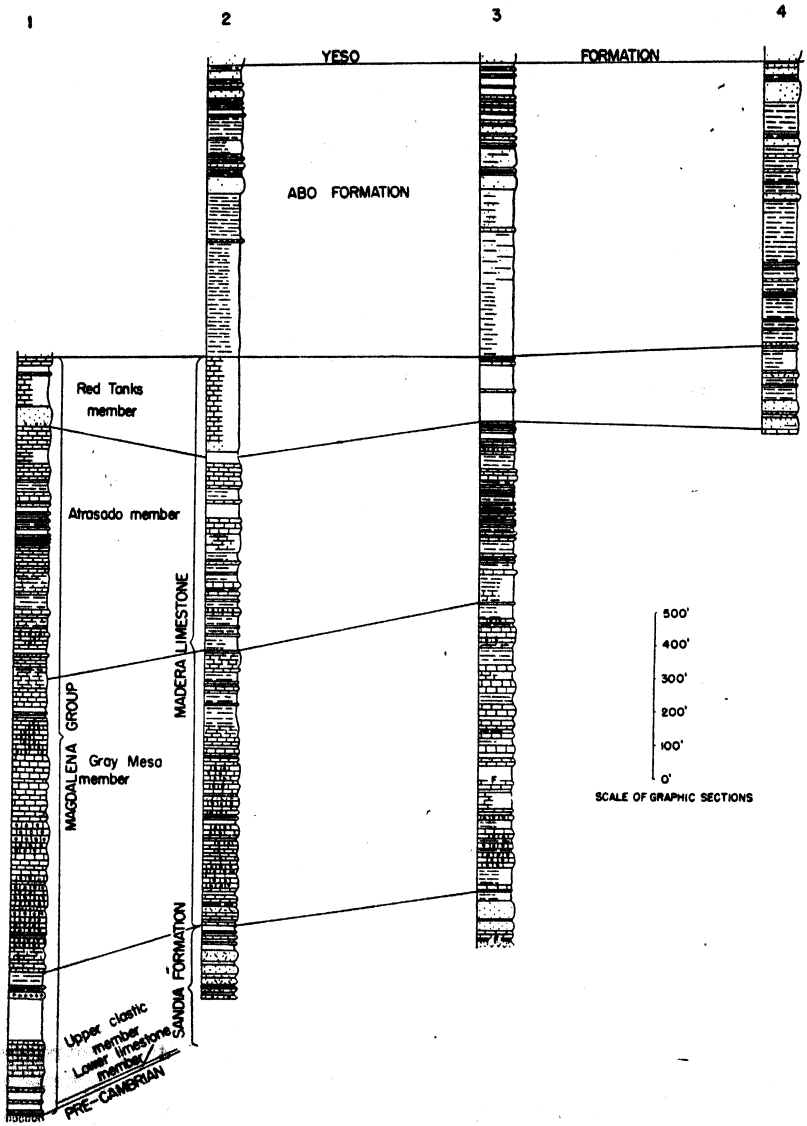
STRATIGRAPHIC SECTIONS

The following group of sections illustrates some of the lithologic variations in the Pennsylvanian series of northern New Mexico. For more detailed data the reader is referred to reports and maps cited above. The purpose of these graphic sections is to give general impressions of typical sequences, and a sufficient number of groups shows the numerous details of correlation to be beyond the scope of this paper.

LUCERO UPLIFT, VALENCIA, SOCORRO, AND BERNALILLO COUNTIES, NEW MEXICO

In the Lucero Uplift, which locally constitutes the margin of the Colorado Plateaus, approximately 2,300 feet of Pennsylvanian sediments, constituting the Magdalena group, are exposed (see Fig. 3). The basal portion of the sequence

[11] Although the clastic-calcareous ratio along the Zuni and Defiance uplift is sufficiently high to suggest that the axial portions of these features may have been major sources of clastic sediments, it should be realized that in such areas only uppermost Pennsylvanian strata are known at or near outcrop belts. In contrast, in the ancestral Rocky Mountain area, far larger fractions of Pennsylvanian time are represented by sediments adjacent to the margins of basins.



GRAPHIC SECTIONS OF PALEOZOIC STRATA FROM NAVAJO GAP TO CARRIZO ARROYO

Numbers of sections are those of localities in list

HORIZONTAL SCALE

0 1 2 3 4 5 6 MILES

FIG. 3

is the Sandia formation, which contains *Fusulinella* spp., and rests on problematical Mississippian strata[12] or directly on pre-Cambrian granite.

Overlying the Sandia formation is the Madera limestone, which has been divided into three local members. The oldest of these is the Gray Mesa member, which consists of an almost continuous sequence of dark-gray limestone containing microfaunas at its base, characterized by the genus *Fusulinella*. A sequence of faunas higher in the member is notable in that it includes a rather complete series of forms of the genus *Fusulina*. Evidently on the basis of microfaunas these latter strata constitute the major portion of the Des Moines group, although the basal portion is possibly of Lampasas age.

Resting conformably on the Gray Mesa member of the Madera limestone is the Atrasado member, an alternation of limestone; gray, olive, and red shale; and arkosic sandstone. At its base a small fauna, including advanced types of *Fusulina* spp. and also *Prismopora* sp., has been identified. Less than 20 feet higher and in the same sequence of thin alternating limestones and gray shale, species of *Triticites* belonging to the *irregularis* group have been identified. Higher still is encountered the standard sequence of species of *Triticites* that are characteristic of Pennsylvanian successions in New Mexico. The zone of *T. ventricosus* and associated forms occurs in the upper part, and therefore the higher beds of this member are considered to be generally equivalent to the Virgil group of the Mid-Continent region.

The youngest strata of the Madera limestone have been separated as a third unit, the Red Tanks member. This consists of red shale and sandstone, capped by one or more ledges of gray limestone containing an abundance of *T. ventricosus*. It may be of Upper Virgil age; but, because of the occurrence of a similar unit a short distance to the southeast containing occasional *Schwagerina* sp. associated with a similar abundance of *T. ventricosus*, the Red Tanks member possibly may be of Permian(?) age.

The Abo formation of Permian(?) age succeeds the Red Tanks member of the Madera limestone with apparent conformity. The nature of the sequence in the upper part of the Madera suggests a gradual transition from marine to continental strata, and no important hiatus appears to separate the Abo from the Madera limestone in this area.

LA JOYA, LOS PIÑOS MOUNTAINS, AND NORTHERN CHUPADERA MESA, SOCORRO AND VALENCIA COUNTIES, NEW MEXICO

The Pennsylvanian strata in this area are included in the Magdalena group and consist of gray cherty limestone, sandstone, shale, and minor amounts of conglomerate (see Fig. 4). These rocks rest with angular unconformity or disconformity on various pre-Cambrian metamorphic and igneous rocks.

The Sandia formation includes gray sandstone, quartz conglomerate, shale, and minor quantities of gray arenaceous limestone. It contains numerous fossils of Lower Pennsylvanian age, including several species of *Fusulinella*.

The Madera limestone conformably overlies the Sandia formation and consists of a lower gray limestone member and an upper arkosic limestone member. These units are approximately equivalent to the Gray Mesa and Atrasado members in the Lucero Uplift. However,

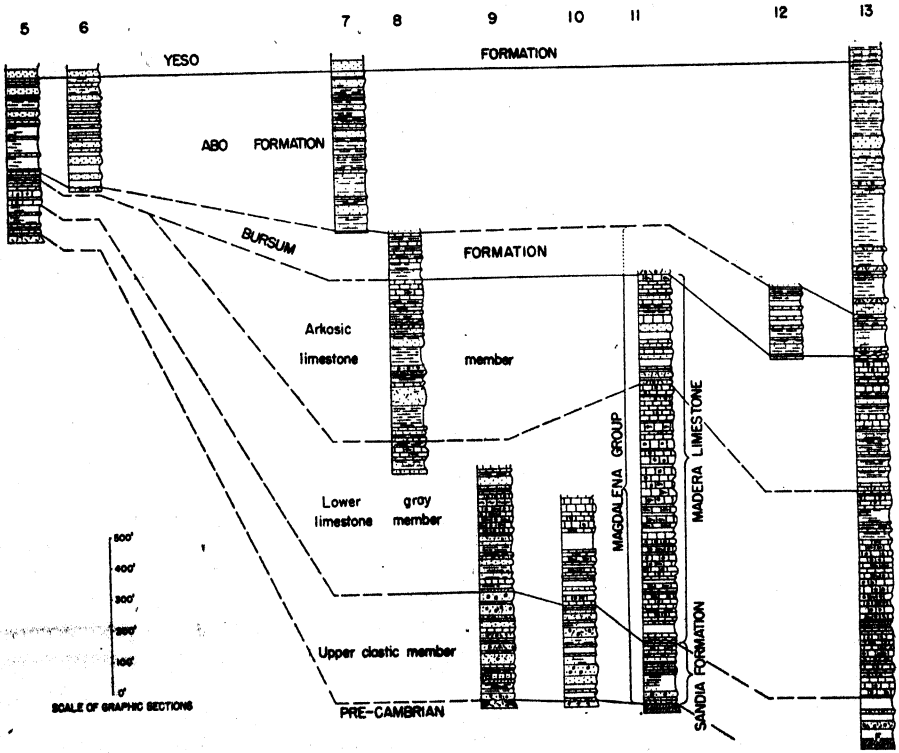
[12] Limestone locally present at the base of the section may be Lake Valley in age. Neither lithologic character nor fossil content supports the correlation sufficiently to permit an unqualified statement.

it is not clear whether the Red Tanks member of the Madera limestone in the Lucero Uplift area is correlative with the upper part of the arkosic limestone member or with the succeeding Bursum formation.

The lower gray limestone member contains massive or thin-bedded cherty limestone interbedded with minor amounts of shale and forms a great cliff in many places.

In contrast, the arkosic limestone member consists of medium- or thin-bedded limestone interbedded with gray, black, and red shale, sandstone, and arkosic sandstone. Conglomerate is usually present. The lower portion of the member is a slope-forming, dominantly clastic unit, whereas the upper portion is usually a cliff-forming sequence of limestone ledges.

Overlying the arkosic limestone member of the Madera limestone in the typical succession is the arkosic and calcareous Bursum formation of Permian(?) age. The latter is succeeded by the Abo formation, which, in this area, consists of arkosic and quartzitic clastic material.



GRAPHIC SECTIONS OF PALEOZOIC STRATA IN LA JOYA, LOS PIÑOS MOUNTAINS, AND CHUPADERA MESA AREAS

Numbers of sections are those of localities in list.

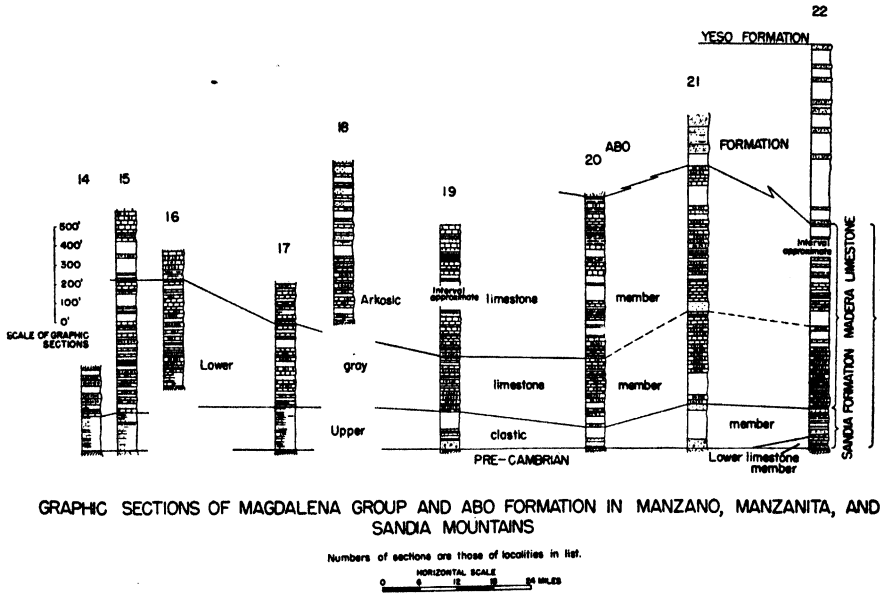
HORIZONTAL SCALE 0 1 2 3 4 MILES

FIG. 4

The lower gray limestone member of the Madera contains marine fossils, including a suite of species of *Fusulina*, indicating a dominantly Des Moines age for the unit. Species of *Fusulinella* occur in abundance, however, in the basal beds of this unit at some localities. In the arkosic limestone member of the Madera, species of *Triticites* are found a short distance above the base of the unit, and a rather complete sequence of forms is known to be present. The Bursum formation contains *Schwagerina* sp. in association with abundant *T. ventricosus*. The thinning of the Magdalena group in the vicinity of the Joyita Hills is of interest. In this area the arkosic limestone member of the Madera is absent, owing to nondeposition on the Joyita axis, a positive feature that appeared during Upper Pennsylvanian time.

RED CANYON, MANZANO MOUNTAINS, TO GALLEGOS RANCH, SANDIA MOUNTAINS

The group of sections shown in Figure 5 extends from Red Canyon, near the southern end of the Manzano Mountains, northward along the eastern slope of the Manzano, Manzanita, and Sandia ranges to Gallegos Ranch, near the northern termination of the Sandia Mountains. As in the other groups of the section, Pennsylvanian marine strata are assigned to the Magdalena group.



GRAPHIC SECTIONS OF MAGDALENA GROUP AND ABO FORMATION IN MANZANO, MANZANITA, AND SANDIA MOUNTAINS

Numbers of sections are those of localities in list.

HORIZONTAL SCALE

FIG. 5

A massive limestone associated with basal clastics occurs in the northern part of the Sandia Mountains below the Sandia formation. It was originally assigned to the Pennsylvanian, but, because it contains no diagnostic fossils and on the basis of recent work in other portions of the state, it is considered by us

to be a possible equivalent of the Leadville limestone, of Mississippian age.

Resting with pronounced disconformity on this limestone of supposed Mississippian age, or directly on the pre-Cambrian granite and metamorphic rocks of the Sandia and Manzano mountains, is the Sandia formation. This unit shows much variation in detail in the several sections illustrated but maintains the rather constant characteristic of alternations of irregularly bedded sandstone, thin coal seams, carbonaceous shale, and occasional impure beds of limestone. Species of *Fusulinella* have been collected from the upper part of the unit in most of the sections.

In the succeeding Madera limestone the lower part of the gray limestone member also carries faunas characterized by species of *Fusulinella*, and the higher parts contain a rather full succession of the several species of this genus. The most advanced species extend into the basal part of the superjacent arkosic limestone member and are there succeeded by a series of species of *Triticites* that culminates in *T. ventricosus*.

Clearly, there is much variation in detail in this group of sections but little variation in general character. There is no physical evidence of any major hiatus in the sequence, which is considered a thin, but rather continuous, representation of marine deposition in a foreland area.

The Abo formation of Permian(?) age overlies the Madera formation with apparent conformity. At most places the contact suggests gradational or transitional relationships.

PEÑASCO ARROYO TO JEMEZ SPRINGS, NACIMIENTO MOUNTAINS, SANDOVAL COUNTY, NEW MEXICO

Guadalupe Box is in the middle of the Rio de las Vacas drainage basin in the heart of the Nacimiento Mountains and offers the only complete and structurally simple section in the range. In the canyon of the Vacas, Pennsylvanian strata rest on a thin representative of the limestone member of the Sandia formation that the writers regard, on lithologic grounds, as a possible equivalent of the Leadville limestone (see Fig. 6).

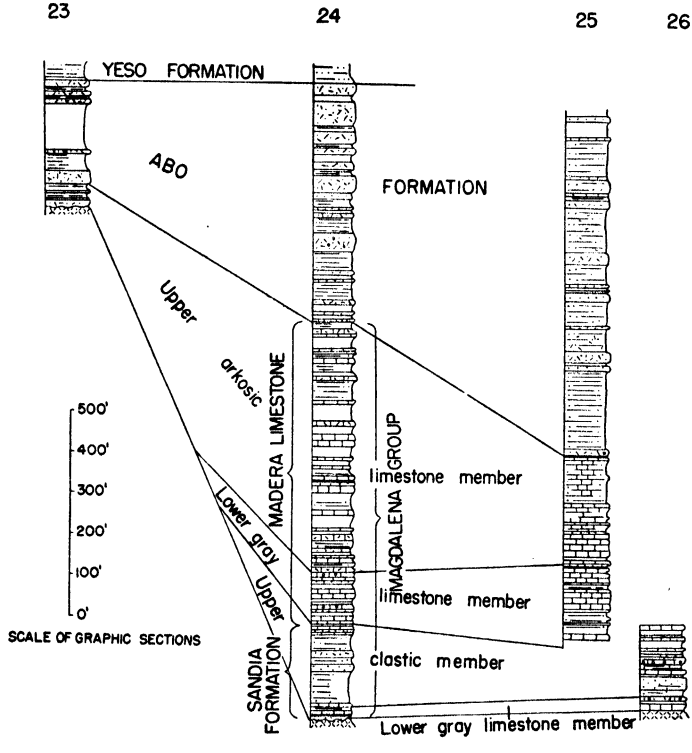
Disconformably overlying this limestone is the upper member of the Sandia formation, which consists of clastic beds and thin limestones. From the basal beds of this member the following fauna has been collected: *Aulopora* sp. cf. *A. prosseri*, *Composita* sp. cf. *C. gibbosa*, *Derbya* sp., *Dictyoclostus* sp. cf. *D. hermosanus*, *Hustedia* sp. cf. *H. miseri*, *Schizophora oklahomae*, *Spirifer occidentalis*. This fauna is believed to be of Morrow age. From the higher parts of the Sandia formation here and at near-by localities, *Millerella* sp., *Prismopora* sp., and *Fusulinella* sp. have been identified. Clearly this part of the Sandia formation in the area is of Morrow and Lampasas age.

In the overlying Madera limestone the lower gray limestone member, which rests conformably on the Sandia formation, has yielded the following fossils: *Chonetina?* sp., *Derbyia crassa*, *Linoproductus* sp. *aff. L. platyumbonus*, *Marginifera* sp. cf. *M. missouriensis*, *Spirifer occidentalis*, *S. rockymontanus*, *Fusulina* spp., *Wedekindelina* sp., *Fusulinella* sp. This fauna indicates that the containing beds are of late Lampasas and of early and middle Des Moines age. We do not recognize, at this time, any hiatuses within the sequence.

The succeeding arkosic limestone member of the Madera contains, immediately above the contact with the lower gray limestone, abundant *Chaetetes milleporaceous;* and 60 feet above the base the following fauna has been found: *Composita trilobata*, *Dictyoclostus colora-*

doensis, *Dielasma bovidens*, *Linoproductus platyumbonus*, *Neospirifer triplicatus*, *Phricodothyris perplexa*, *Punctospirifer kentuckyensis*, *Acanthopecten carboniferus*, *Solenomya?* sp. Approximately 280 feet above the base of the member the following forms have been collected: *Neospirifer triplicatus*, *Allorisma terminale*, *Edmondia* sp., *Pseudomonotis equistriata*, *Bellerophon* sp. cf. *B. giganteus*, *Straparollus* (*Euomphalus?*) sp. Near the top of the member, *N. triplicatus*, *Marginifera wabashensis*, *A. terminale*, and *T. ventricosus* have been found. This sequence is rather typical of the lithologic and faunal succession in the upper member of the Madera limestone. There appear to be no important disconformities, and probably the sequence represents slow but nearly continuous sedimentation.

The group of sections in Figure 6 shows interesting variations in passing westward to Peñasco Arroyo. Clearly, the Sandia formation and the lower member of the Madera limestone dis-



GRAPHIC SECTIONS OF PALEOZOIC STRATA FROM PENASCO ARROYO TO JEMEZ SPRINGS, NEW MEXICO

Numbers of sections are those of localities in list.

HORIZONTAL SCALE

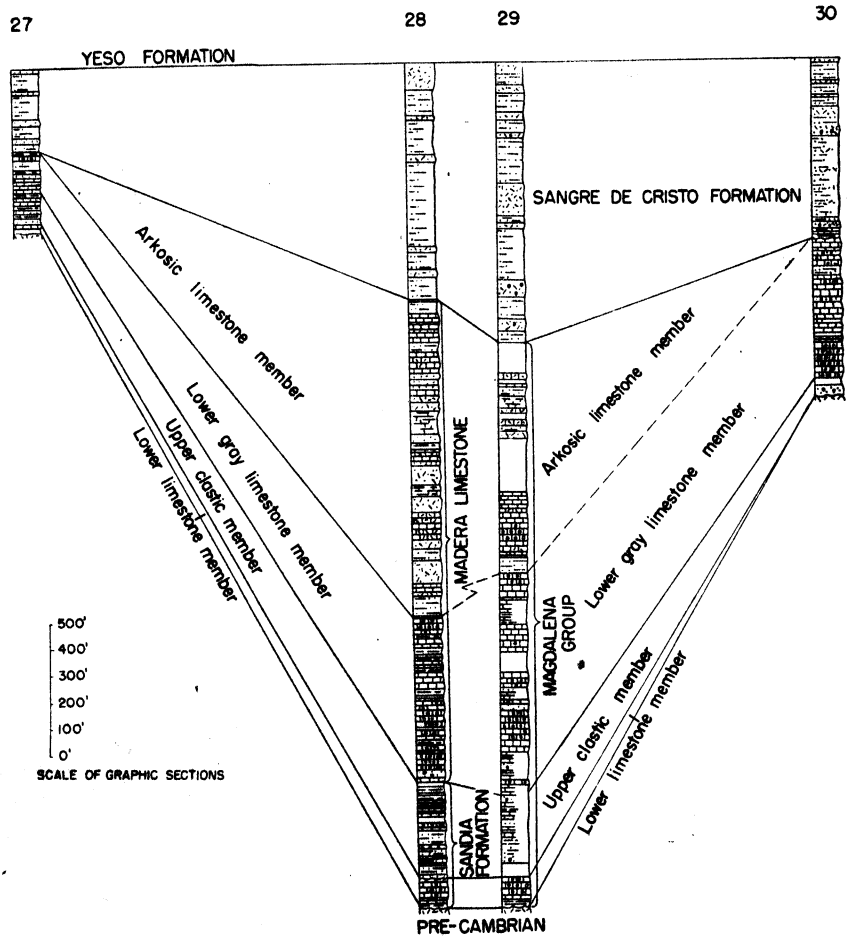
0 1 2 3 4 6 8 MILES

FIG. 6

appear, and the upper arkosic limestone member of the Madera thins. The absence of the lower part of the Magdalena group may be explained by overlap, and the thinning of the upper unit of the group probably resulted from the rise of a strong, positive element during late Pennsylvanian time.

LAMY TO BERNAL, SANGRE DE CRISTO MOUNTAINS, SANTE FE AND SAN MIGUEL COUNTIES, NEW MEXICO

Along Pecos Valley in the southernmost part of the Sangre de Cristo Mountains, Pennsylvanian strata referred to the Magdalena group are well exposed (see Fig. 7).



GRAPHIC SECTIONS OF PALEOZOIC STRATA IN SANGRE DE CRISTO MOUNTAINS

Numbers of sections are those of localities in list.

HORIZONTAL SCALE
0 6 12 18 24 MILES

FIG. 7

Resting on the pre-Cambrian or on a regolith, massive limestone has been classified tentatively as the lower member of the Sandia formation. This limestone may be, however, of Mississippian age or older, but it has yielded no diagnostic fossils. Overlying it is the upper, or clastic, member of the Sandia formation, which consists of shale, sandstone, impure limestone, and thin coal. Both the invertebrate and the fossil plants of these beds suggest correlation with strata of Lampasas age elsewhere.

The Madera limestone, which conformably succeeds the Sandia formation, consists of a lower, gray limestone member and an upper, arkosic limestone member. The lower member is dominantly limestone, with minor alternations of calcareous shale and small quantities of sandstone. It contains, in its lower portion, several species of *Fusulinella* and in its upper portion species of *Fusulina*. Probably this member is chiefly of Des Moines age, although possibly the lowermost strata are of Lampasas age.

Overlying the lower gray limestone member of the Madera limestone, and at many points transitional with it, are alternations of limestone, sandy and arkosic limestone, arkose, and red, olive, and gray shale that constitute the arkosic limestone member. In the lowest beds of this unit, fossils of Upper Des Moines age have been collected. Numerous invertebrates in the succeeding portions of the member suggest, however, an Upper Pennsylvanian age assignment, as do also fossil plants collected in its middle and upper parts.

The Sangre de Cristo formation, a highly variable succession of arkosic and red beds, conformably or unconformably overlies the Magdalena group. Possibly the lower portion of this unit is, at some localities, Pennsylvanian in age. The middle and upper parts have been dated, on the basis of fossil plants, as Permian(?).

Both east and west of Pecos Valley the Pennsylvanian rocks thin, because of the development of late Pennsylvanian positive elements. At Lamy, New Mexico, a thin sequence of the Sandia formation and the lower member of the Madera are overlain by arkose of the Sangre de Cristo formation. A similar occurrence is apparent in the vicinity of Bernal.

LIST OF LOCALITIES OF GRAPHIC SECTIONS

1. Ladron Mountains: sections 9, 10, 14, 15, and 23, T. 3 N., R. 3 W., Socorro County, New Mexico.
2. Monte de Belen: sections 13, 14, 23, 24, and 32, T. 4 N., R. 3 W., Socorro County, New Mexico.
3. Gray Mesa: section 13, T. 5 N., R. 3 W., and section 25, T. 6 N., R. 3 W., Socorro County, New Mexico.
4. Carrizo Arroyo: section 7, T. 6 N., R. 2 W., and section 1, T. 6 N., R. 3 W., Valencia County, New Mexico.
5. Central part of Joyita Hills, Sevilleta Grant, Socorro County, New Mexico.
6. Northern part of Joyita Hills, Sevilleta Grant, Socorro County, New Mexico.
7. Headwaters of Arroyo Cibola, southern Los Piños Mountains, Sevilleta Grant, Socorro County, New Mexico.
8. Head of Arroyo Cibola, Los Piños Mountains, Sevilleta Grant, Socorro County, New Mexico.
9. South-central part of Los Piños Mountains, Sevilleta Grant, Socorro County, New Mexico.
10. Central part of Los Piños Mountains, Sevilleta Grant, Socorro County, New Mexico.
11. East-central part of Los Piños Mountains, Sevilleta Grant, Socorro County, New Mexico.
12. Montosa Arroyo: sections 13 and 14, T. 2 N., R. 4 E., Socorro County, New Mexico.
13. Abo Canyon: southeastern part of Casa Colorado Grant, and portions of T. 2 and 3 N., R. 5 and 6 E., between Sais Quarry

and Abo State Monument, Socorro and Torrance Counties, New Mexico.

14. Red Canyon: section 28, T. 5 N., R. 5 E., Torrance County, New Mexico.
15. Arroyo de los Piños Reales: section 21, T. 5 N., R. 5 E., Torrance County, New Mexico.
16. New Canyon: sections 10 and 11, T. 5 N., R. 5 E., Torrance County, New Mexico.
17. Burro Mountain: section 14, T. 6 N., R. 5 E., Torrance County, New Mexico.
18. Tajique Arroyo: west of Tajique, Torrance County, New Mexico.
19. Hell Canyon: sections 21 and 22, T. 8 N., R. 5 E., Bernalillo County, New Mexico.
20. Cedro Canyon: sections 26 and 35, T. 10 N., R. 5 E., Bernalillo County, New Mexico.
21. San Antonito vicinity: section 22, T. 11 N., R. 6 E., Bernalillo County, New Mexico.
22. Gallegos Ranch: 1½ miles east of Placitas, Sandoval County, New Mexico.
23. Peñasco Arroyo: sections 20 and 21, T. 16 N., R. 1 E., Sandoval County, New Mexico.
24. Guadalupe Box: Cañon de San Diego Grant, Sandoval County, New Mexico.
25. Jemez Springs: Cañon de San Diego Grant, Sandoval County, New Mexico.
26. Soda Dam: Cañon de San Diego Grant, Sandoval County, New Mexico.
27. Vicinity of Apache Canyon, near Lamy, Santa Fe County, New Mexico.
28. Pecos River: section 33, T. 17 N., R. 12 E.; and sections 4, 9, 16, and 20, T. 16 N., R. 12 E., and in the Pecos Grant, San Miguel County, New Mexico.
29. Vallecitos: vicinity of Cow Creek, sections 7 and 18, T. 16 N., R. 13 E., San Miguel County, New Mexico.
30. Bernal Creek, four miles north of Bernal, San Miguel County, New Mexico.

REGIONAL VARIATIONS IN PENNSYLVANIAN LITHOLOGY

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ABSTRACT

Recurring similar sequences of strata characterize the Pennsylvanian in most regions in the United States. The Illinois type of sequence has been called the "cyclothem" and the Kansas type the "megacyclothem." There are many variations in the pattern of these rhythmic successions, and within the Pennsylvanian areas of the United States they include representatives of nearly all kinds of sedimentary rocks. Thirty-six representative Pennsylvanian successions, two from the Mississippian, and one from the Lower Permian are graphically represented. The rhythmic successions may record diastrophic or climatic fluctuations of continent-wide extent, and the varied type of successions composing them may result from locally variable features of environment, topography, climate, or the configuration of land and sea.

Data sheets to record quantitative lithologic variations in the Pennsylvanian and its principal subdivisions have been devised. From these sheets, prepared on the basis of one for each county, various types of lithologic-variation maps and isopach maps can be prepared.

INTRODUCTION

The arrangement of Pennsylvanian strata of varied lithology into an orderly sequence several times repeated was first described by J. A. Udden from the Peoria quadrangle, Illinois.[1] The more widespread occurrence of such a rhythmic pattern was reported by J. M. Weller.[2] He stated that in western Illinois the succession consists ideally of ten beds, named as follows in ascending order: (*a*) sandstone, (*b*) sandy and micaceous shale, (*c*) "fresh-water" limestone, (*d*) underclay, (*e*) coal, (*f*) gray nonmarine shale, (*g*) marine limestone, (*h*) black slaty shale, (*i*) marine limestone and calcareous shale, and (*j*) shale containing ironstone bands or nodules in the upper part and thin limestone layers in the lower part. Each succession is locally separated from underlying and overlying successions by an erosional unconformity. To this frequently repeated rhythmic succession the name "cyclothem" was applied.[3] A symposium held at Urbana in 1930 brought out the fact that successions of cyclic or rhythmic character were to be found in Ohio,[4] West Virginia,[5] and the northern and southern mid-continent areas,[6] as well as in Illinois; but the details of the rhythmic succession in each of these areas are unlike those of Weller's western Illinois cyclothem. The plane of separation between cyclothems was placed by Weller at the base of the sandstone member (*a*), but Stout prefers the base of the coal (*e*), Reger the top of the sandstone (*a*), and Moore the top of the upper marine limestone (*i*) as more convenient planes of separation in their

[1] "Geology and Mineral Resources of the Peoria Quadrangle," *U.S. Geol. Surv. Bull. 506* (1912), pp. 27, 47–50.

[2] "Cyclical Sedimentation of the Pennsylvanian Period and Its Significance," *Jour. Geol.*, Vol. XXXVIII (1930), pp. 97–135; "The Conception of Cyclical Sedimentation during the Pennsylvanian Period," *Ill. Geol. Surv. Bull. 60* (1931), pp. 163–77.

[3] H. R. Wanless and J. M. Weller, "Correlation and Extent of Pennsylvanian Cyclothems," *Bull. Geol. Soc. Amer.*, Vol. XLIII (1932), p. 1003.

[4] Wilber Stout, "Pennsylvanian Cycles in Ohio," *Ill. Geol. Surv. Bull. 60* (1931), pp. 195–216.

[5] David B. Reger, "Pennsylvanian Cycles in West Virginia," *Ill. Geol. Surv. Bull. 60* (1931), pp. 217–39.

[6] Raymond C. Moore, "Pennsylvanian Cycles in the Northern Mid-continent Region," *Ill. Geol. Surv. Bull. 60* (1931), pp. 247–57; Frederick B. Plummer, "Pennsylvanian Sedimentation in Texas," *Ill. Geol. Surv. Bull. 60* (1931), pp. 259–69.

respective areas if the cyclothem is to be considered primarily a formational unit to be used in geologic mapping. Later, Moore described a megacycle or cycle of cycles from the Kansas section,[7] recording four or five rather than one, marine invasion, but with a different sedimentary record for each until the corresponding cycle of the next megacycle has been reached (Fig. 3, col. 4). A form of cyclic sedimentation in the Rocky Mountains has been described[8] and one each from Michigan,[9] England,[10] the Silesian Basin,[11] and the Donetz Basin.[12] During the summer of 1934 the author spent a few weeks studying the Pennsylvanian succession in various parts of Colorado, New Mexico, Wyoming, South Dakota, and Utah. He found that at virtually every place the succession of rock types exhibits some form of rhythmic pattern but that the rock sequence recording the rhythm varies greatly from place to place and nowhere greatly resembles Weller's western Illinois cycle, Stout's Ohio cycle, or Moore's Kansas megacycle. The Rocky Mountain successions differ from those farther east in that they contain very little carbonaceous sediment (coal or black shale) and comparatively little argillaceous sediment (shale and underclay) and in that the sandstones are more generally arkosic than in areas farther east. The sequence composing the rhythmic alternation is generally simpler, being composed of two or three rock types rather than eight or ten, as in the areas farther east. These western successions are reported to be much less widespread than those farther east,[13] and they may not record comparable periods of time. Since 1934, opportunity has been afforded the author to study the Pennsylvanian successions in Indiana, Ohio, Kentucky, Pennsylvania, West Virginia, Virginia, Tennessee, and Georgia during 1937–39 and Wyoming during 1945; and brief field excursions to Pennsylvanian areas in Alabama, Iowa, Missouri, Oklahoma, Texas, and New Mexico have provided additional illustrations of variation in cyclic successions.

In this paper it is proposed to present some of the varied types of succession from the Pennsylvanian of the United States that may be considered cyclic. There are also shown two columns (Fig. 2, col. 8, and Fig. 4, col. 1) illustrating cyclic successions from the later Mississippian (Chester and Brazer) and one (Fig. 3, col. 9) from the Early Permian (Big Blue), showing that types of rhythmic successions are to be found in systems adjoining the Pennsylvanian. The suggestion is made that the various sequences are somewhat comparable in the time required for their deposition and that they result from changes in sea-level, changes in climate, and diastrophic disturbances not entirely local to the

[7] "Stratigraphic Classification of the Pennsylvanian Rock of Kansas," *Kan. Univ. Bull. 22* (1935), pp. 29–35.

[8] J. Harlan Johnson, "Contribution to the Geology of the Sangre de Cristo Mountains of Colorado," *Proc. Colo. Sci. Soc.*, Vol. XII (1929), pp. 1–11.

[9] W. A. Kelly, "Pennsylvanian Stratigraphy near Grand Ledge, Michigan," *Jour. Geol.*, Vol. XLI (1933), pp. 79–86.

[10] R. G. Hudson, "On the Rhythmic Succession of the Yoredale Series in Wensleydale," *Proc. Yorkshire Geol. Soc.*, Vol. XX, Part I (1923–24), pp. 1–11; James Brough, "On Rhythmic Deposition in the Yoredale Series," *Proc. Univ. Durham Philos. Soc.*, Vol. VIII, Part II (1929), pp. 116–25.

[11] Axel Born, "Periodizität epirogener Krustenbewegungen," *Rept. 16th Internat. Geol. Cong., 1933*, Vol. I (1936), p. 172.

[12] B. Tschernyschew, "Carbonicola, Anthracomya, and Naiadites of the Donetz Basin," *Trans. Geol. and Prospect. Serv., U.S.S.R.*, Fasc. 72 (1931), pp. 109–11; A. W. Grabau, "The Permian of Mongolia," *Amer. Mus. Nat. Hist., Natural History of Central Asia*, Vol. IV (1931), pp. 432–42.

[13] C. B. Read, personal communication.

areas represented. Correlation is not intended between the various columns, as the beds in them represent all the major groups of the Pennsylvanian. The columns have been drawn with the base of the coal at the same level near the middle of the column where coal is present. The assumed position of a coal horizon has been used to line up those columns which do not include coal beds. The columns are intended to illustrate how numerous kinds of rocks fit into a broadly generalized cyclic pattern. Each column is based on a particular outcrop section measured by the author or derived from the published literature. No effort was made to generalize or idealize the sequence for particular localities. The distribution of marine and nonmarine invertebrates and fossil plants, where known, is indicated by symbols.

TYPES OF PENNSYLVANIAN CYCLES IN THE APPALACHIAN COAL FIELD

The Appalachian Pennsylvanian is characterized by a predominance of clastic sediments, largely nonmarine; by an abundance of coal beds; and, in restricted areas, by refractory fire clays, flints, sedimentary iron ores, red beds, and fresh-water limestones. The cyclic successions range from more than 100 feet thick (Fig. 1, cols. 8 and 9) near the southeast border of the coal field to 25 feet or less (Fig. 1, col. 4) in Ohio and part of western Pennsylvania.

In Figure 1, column 1 illustrates the development of refractory fire clay in the position of the underclay (*d*) of Weller's Illinois cycle. The refractory clay may be of flint, semiflint, or plastic type. Refractory clays seem to be restricted to localities in which a part of the Pennsylvanian is thinner than normal, illustrating either very slow sedimentation or prolonged weathering between brief periods of sedimentation. The section shows the Sciotoville clay of southern Ohio of Pottsville age; but other refractory clays occur higher in the Pottsville and in the Allegheny in Ohio, northern West Virginia, Maryland, and Pennsylvania. The origin of these clays has been discussed by Wilber Stout[14] and others.

Column 2 (Fig. 1) illustrates the occurrence of bedded dark chert or flint in cyclic succession. The example shown is the Kanawha black flint of West Virginia, of Kanawha age. Other examples are the Flint Ridge flint of Kentucky and the Upper Mercer, Zaleski, Vanport, and Brush Creek flints of Ohio, of Pottsville, Allegheny, and Conemaugh ages. The flint yields a marine fauna, seems to occupy the position of the upper marine limestone (*i*) of Weller's cyclothem, and grades laterally into marine limestone (Vanport and Upper Mercer) or into black slaty shale (Kanawha). These flints are restricted in distribution in the Appalachian region but were extensively quarried by Indians as material for artifact construction, and they are useful marker beds in stratigraphic work. This column also illustrates what is called a "rider" coal, a thin additional coal above the main coal bed of a cycle, which apparently belongs in the same cycle with it.

Column 3 (Fig. 1) illustrates the thin "black-band" iron ores characteristic of the Pottsville and Lower Allegheny of the old Hanging Rock iron district of southern Ohio. The cycle shown includes the Lower Mercer limestone and Middle Mercer coal and indicates that the iron-ore bands belong above the main marine limestone of the cycle. They seem to correspond with the ironstone nodules and thin bands of the upper shale (*j*)

[14] "Origin of Coal Formation Clays," *Ohio Geol. Surv. Bull. 26* (4th ser., 1923), pp. 533–68.

of Weller's cyclothem. The ironstone beds ("black-band" ore) and nodules ("kidney" ore) seem to be somewhat commoner in dark-gray to black carbonaceous shales than in lighter shales. Some of the iron-ore bands contain marine fossils and other brackish or fresh-water invertebrates. Column 7 shows several levels of nodules and four thin bands in dark shale of the Briceville formation of Tennessee.

Column 4 (Fig. 1) illustrates the "fresh-water" limestones, which are perhaps the most conspicuous members of Upper Conemaugh, Monongahela, and Dunkard cycles in parts of east-central Ohio, northern West Virginia, and southwestern Pennsylvania. They seem to occupy the position of Weller's "fresh-water" limestone (*c*) beneath, rather than above, the coal beds. They differ, in that Weller's limestone is largely unfossiliferous, whereas the Appalachian fresh-water limestones contain a rich fauna of gastropods, pelecypods, ostracodes, and fish remains, all apparently of fresh-water facies. This facies has a prominent development near Washington, Pennsylvania, and Wheeling, West Virginia; but the limestones thin and disappear rather abruptly a little south of that area, along with the coal beds and plant-bearing shales, the succession giving way to red beds, as illustrated in column 5. This facies change affects later Conemaugh, Monongahela, and Dunkard beds. In the example in column 5 the Ames limestone and Harlem (?) coal are present near Charleston, West Virginia; but a little farther east the limestone becomes nonmarine, the coal thins out and disappears, and the cyclothem becomes almost entirely red shale and sandstone. At their northern margin the red beds seem to replace the sandy shale (*b*) of Weller's cycle, but the upper shale (*j*) and the sandstone (*a*) may also grade into red sediments.

Column 6 (Fig. 1) illustrates the Charleston sandstone near Charleston, West Virginia, which is formed by the union of the Roaring Creek, East Lynn, Freeport, Mahoning, and Buffalo sandstones of Kanawha, Allegheny, and Lower Conemaugh ages, the other strata of the normal succession largely disappearing. The composite character of the sandstone near Charleston may be recognized in that it is interrupted at about 50-foot intervals by thin but persistent coal beds and by lithologic differences between the sandstones above and below a particular coal. The No. 5 Block, Upper Kittanning, Lower Freeport, Upper Freeport, and Mahoning coals all seem to fall in this great sandstone mass. Although there are abrupt unconformable contacts at the bases of the several sandstones, there is no certainty that shales were deposited above each coal and subsequently eroded. The Charleston region may have been part of a large area of alluvial-fan or delta accumulation of sand, interrupted only briefly by coal-forming conditions throughout the later Kanawha, Allegheny, and early Conemaugh. Here the equivalents of Weller's beds (*a*) and (*e*) alone record the cycle.

Column 7 (Fig. 1) was chosen to illustrate thick, dark-gray to black upper shales yielding no fossils or occasional fresh-water pelecypods or brackish-water *Lingula* faunas. Plant remains are uncommon or absent. Although the dark shales are probably the counterparts of Weller's upper shale (*j*) and resemble them in containing ironstone nodules and bands, they were evidently deposited under conditions less favorable to the oxidation of organic matter. The scarcity of fossils in the dark shales suggests that the water may have been toxic to many

kinds of life-forms. The example is from the middle part of the Briceville formation of Tennessee above the Poplar Creek coal; but equally prominent dark shales occur in the Eastland shale of Tennessee, older than the Briceville; the Whetstone Creek shale and the shale above the Williamsburg coal of Kentucky; the Eagle shale; shales associated with the Douglas and Iaeger coals of southern West Virginia; and shales associated with the Quakertown coal of Ohio and Pennsylvania. Their period of greatest prominence seems to correspond with the Lampasas stage.

Column 8 (Fig. 1) illustrates a type of cyclic succession displayed in the Wise formation of Virginia; the Hance, Mingo, and Hignite formations of the Cumberland Gap coal field; and the Scott and Anderson formations of Tennessee. Here at the southeastern border of the coal field a thick, massive, basal sandstone is separated from the next similar bed by a shaly zone, generally containing two to ten benches of coal separated by shale, underclay, or even sandstone. Rapid sedimentation is attested by the finding of erect fossil casts of tree trunks rising several feet across shales or sandstones, in some places rooted in the top of a coal bed. The coal may be very thick and locally unparted, like the famous Pocahontas No. 3 bed at Pocahontas, Virginia, and may include zones of cannel coal. The outcropping succession in Wise County, Virginia, displays forty-four cycles of this type in beds of Lampasas age alone, with an average thickness of 68 feet, in contrast with Fulton County, Illinois, where the strata of Lampasas age display four cycles, averaging 10 feet.

Column 9 (Fig. 1) is representative of the Lee (Morrowan) of northern Tennessee and parts of Kentucky, Georgia, and Virginia, with a thick and locally conglomeratic basal sandstone overlain by a shaly coal-bearing, easily eroded zone. The example shows the Sewanee coal and conglomerate of southern Tennessee. Such a cycle as this includes two formations in the southern Tennessee system of nomenclature. The example drawn includes the Sewanee conglomerate and Whitwell shale. Overlying pairs of formations composing a cycle are (1) Herbert sandstone and Eastland shale, (2) Newton sandstone and Vandever shale, and (3) Rockcastle sandstone and Duskin Creek shale. The outcrops of this type of cycle give rise to plateaus, benches, or hills of circumdenudation bordered by talus-covered forested slopes on the shale outcrop and lower benches on the next lower sandstones.

TYPES OF PENNSYLVANIAN CYCLES IN THE MICHIGAN AND EASTERN INTERIOR BASINS

The Pennsylvanian rocks of the Michigan Basin (Fig. 2) are generally deeply covered with glacial till, and only one small area of outcrop seems to have had careful examination. This area, near Grand Ledge,[15] displays a section about 130 feet thick which Kelly divided into eight cyclical formations. The strata seem to be of Lampasas or Lower Des Moines age or both, judging from the fauna listed. Although there are individual differences in sequence in the eight cycles, the column drawn of cyclical formation *F*, the sixth from the base, shows general similarity to the sequence in western Illinois described by Weller, whose zones (*a*), (*b*), (*d*), (*e*), (*f*), (*h*), (*i*), and (*j*) are apparently all represented but by lesser thicknesses than are generally found in Illinois.

[15] W. A. Kelly, "Pennsylvanian Stratigraphy near Grand Ledge, Michigan," *Jour. Geol.*, Vol. XLI (1933), pp. 79–86.

In Figure 2, column 2 represents a type of cyclothem characteristic of the McLeansboro of northern Illinois. The example shown includes the La Salle limestone, the thickest marine Pennsylvanian limestone of northern Illinois. Other cyclothems—the Little Vermillion above the La Salle and the Hall, Hicks, and Turner below—differ principally in having a thinner limestone, averaging about 2 feet instead of 25 feet thick. This group of cyclothems lacks a basal sandstone and a fresh-water limestone and has thin coal and thin, largely calcareous underclay. It consists mostly of beds (*g*), (*h*), (*i*), and (*j*) of Weller's cyclothem.

Column 3 (Fig. 2), which represents the Trivoli cyclothem and includes coal No. 8, Lower McLeansboro, of western Illinois, is quite representative of Weller's typical cyclothem. Bed (*b*), sandy shale, is absent or very thin; bed (*f*), nonmarine shale, is absent; and the "fresh-water" limestone (*c*) is represented by somewhat infrequent limestone nodules in calcareous clay.

Column 4 (Fig. 2), the Liverpool cyclothem, including the Colchester (No. 2) coal of the Lower Carbondale, western Illinois, resembles the preceding in most respects but illustrates a thick development of the nonmarine shale (*f*), with no marine limestone in the position of Weller's bed (*g*), i.e., between the black slaty shale (*h*) and the coal. The black slaty shale may contain large, oval, cannon-ball or irregularly shaped concretions, locally called "niggerheads." The upper marine zone (*i*) is here characterized by an alternation of several thin bands of limestone and shale, the Oak Grove member, each with distinctive lithology and fauna, rather than by a single bed of limestone. The nonmarine shale (*f*) shows marked local variability, ranging from 20 feet to a few inches within less than a mile. It is equivalent to the stratum containing concretions in northern Illinois, which yield the well-known Mazon Creek flora.

Column 5 (Fig. 2) is typical of the Carbondale and Lower McLeansboro of the Ozark flank region of western and southwestern Illinois. The column illustrated is the Brereton cyclothem, in Brown County, which here consists of underclay (*d*), a sooty film representing the Herrin (No. 6) coal, which is 4–11 feet thick elsewhere in Illinois, and the marine limestone (*h*). The underclay of the next cyclothem overlies the Brereton limestone; and the marine St. David limestone, cap of the Springfield (No. 5) coal, is below the underclay of the No. 6 coal. Here only 5 or 6 feet record a widespread cyclothem, which is generally 30–60 feet thick. This condition obtains in Adams, Brown, and Calhoun counties west of the Illinois River and in Greene, Jersey, Madison, St. Clair, and Randolph counties opposite St. Louis, as well as in the Pennsylvanian outlier in and near St. Louis, Missouri.[16] The slight thickness probably results from weak positive tendencies of the Ozark region during Carbondale and McLeansboro time, but the fine texture of the sediment indicates that the Ozark dome was not a source area for much sediment.

Column 6 (Fig. 2) represents a column near Alton, Madison County, Illinois, in the St. Louis region, in which six cyclothems representing Lampasas and early Des Moines time show a composite thickness of only about 40 feet. These much-reduced cyclothems are correlated tentatively, from the base up, with the Baby-

[16] J. Brookes Knight, "The Location and Areal Extent of the St. Louis Pennsylvanian Outlier," *Amer. Jour. Sci.*, Vol. XXV (5th ser., 1933), pp. 25–48, 166–78.

lon, Pope Creek, Seville, De Long, Seahorne, and Liverpool cyclothems. The most conspicuous feature of the succession is refractory fire clay, which is known as the "Cheltenham clay" in the St. Louis region. Thin smut streaks in the clay seem to mark the positions of the Pope Creek, De Long, and Seahorne coals. The slightly thicker Seville coal and the marine Seahorne limestone also wedge out in part of the St. Louis region, the fire clay then extending from the Pope Creek up through the De Long, Seahorne, and Liverpool clays to the base of the No. 2 coal. This association of refractory clay with slow sedimentation has been previously noted in Figure 1, column 1, for the Appalachian region; and similar successions prevail in Lincoln and Audrain counties, Missouri, and in La Salle and Grundy counties, northern Illinois.

Column 7 (Fig. 2) represents a type of cyclothem which is common in the Lower Tradewater and Caseyville of southern Illinois, Indiana, and western Kentucky. The column is the Makanda sandstone and overlying underclay, coal, and shale from Jackson County, Illinois; but several successions above and below this are quite similar. The basal sandstone is thick and may be conglomeratic. The underclay is thin and shaly, the coal is thin and discontinuous, and the overlying shales are nonmarine and may be truncated by the next-overlying sandstone. It somewhat resembles the southern Appalachian Lee cyclothem (Fig. 1, col. 9).

Column 8 (Fig. 2) illustrates cyclical sedimentation in the Chester (later Mississippian) of southern Illinois. This series consists of sixteen formations, eight pairs of sandstone and marine shale-limestone alternations, as follows, from the base up: (1) Aux Vases–Renault, (2) Bethel–Paint Creek, (3) Cypress-Golconda, (4) Hardinsburg–Glen Dean, (5) Tar Springs–Vienna, (6) Waltersburg-Menard, (7) Palestine-Clore, the pair shown in this column, and (8) Degonia-Kinkaid. Coal is rarely more than a few inches thick and less persistent than Pennsylvanian coals, but the succession seems to record eight distinct cyclic alternations, with an average of about 100 feet in each. The sandstones are finer grained and more quartzitic, and the limestones are generally thicker and more fossiliferous, than in the Illinois Pennsylvanian.

TYPES OF PENNSYLVANIAN CYCLES IN THE MID-CONTINENT REGION

In Figure 3, column 1 is somewhat representative of the Upper Des Moines of southern Iowa. The cyclothem shown includes the Mystic coal and corresponds with the Brereton cyclothem of Illinois. The succession differs from that for the same cyclothem in western Illinois, in that the Cuba sandstone of western Illinois has graded to sandy shale, and a marine limestone, the Higginsville, beneath the coal, seems to be the counterpart of the "fresh-water" limestone below the No. 6 coal of Illinois. This change of Weller's bed (*c*) from a "fresh-water" to a marine limestone is quite characteristic of the mid-continent region. Columns 3 and 4 of this figure show other examples of these lower marine limestones.

Column 2 (Fig. 3), in Nebraska, shows that the upper marine limestone (Weller's [*i*]) is greatly expanded and that there is no coal, little underclay, and no basal sandstone. The nonmarine part of the cycle here is red shale in place of sandy and micaceous shale above sandstone. The equivalent of the underclay in this and the succeeding columns can be

recognized as a nonlaminated shale below laminated shale, rather than by the slickensides and root traces which characterize it farther east. The cycle illustrated includes the Iatan limestone at the top of the Missouri group.

Column 3 (Fig. 3) is representative of the Marmaton (Upper Des Moines) type of cycle of Missouri and eastern Kansas.[17] It is simpler than the megacyclothem and somewhat intermediate between it and Weller's Illinois cyclothem, indicating that the megacyclothem probably represents the same magnitude of time as the Illinois cyclothem and not a grouping of the latter. The example shown includes the Altamont limestone in north-central Missouri.

Column 4 (Fig. 3) represents Moore's megacyclothem, typically developed in the Shawnee group of Virgil age, with virtually no coal or underclay, little sandstone, and several zones of marine limestone and shale, which are much more prominent than in Illinois.[18] The example shown includes the Oread limestone and the lower part of the Kanwaka shale above and the upper part of the Lawrence shale below. The five marine limestones shown probably correspond with beds of the Illinois cyclothem as follows: (1) the lowest limestone below the underclay seems equivalent to the "fresh-water" limestone (*c*); (2) the second limestone between the underclay and the black slaty shale is bed (*g*) of Weller's cyclothem and is the most uniform in thickness and characteristics of any member of the Kansas megacyclothem, though rarely found and quite variable in Illinois; (3) above the black slaty shale is the thickest limestone, corresponding with (*i*) of the Illinois cyclothem; (4) a little above, the main marine limestone may be equivalent to the calcareous shale or "clod" above the Illinois marine limestone (*i*); (5) the uppermost limestone is well within the thick upper shale (*j*) and may be equivalent to occasional thin limestones in Illinois, which occur in the upper shale. The St. David cyclothem of Carbondale age has such a limestone in Fulton and Peoria counties, western Illinois.

In column 5 (Fig. 3), in the Marmaton of northern Oklahoma, the limestone is thicker than in Missouri (col. 3), and each limestone tends to cap an eastward-facing escarpment. The coal and underclay are absent, and the lower marine limestone of column 3 may have graded into a sandstone, the upper of the two sandstones shown. The limestone shown is the Pawnee, near Nowata, Oklahoma.

South of the Arkansas River in Oklahoma the section thickens tremendously with the introduction of clastic units and the thinning and disappearance of limestones. Coals are not so frequent as in somewhat comparable clastic successions of West Virginia, Kentucky, and southern Illinois; but there is a coal in the example illustrated in column 6 (Fig. 3), from the Hartshorne sandstone formation. Somewhat similar sequences, but generally with slightly more marine beds and less coal, are found in the McAlester, Savanna, Boggy, Thurman, Stuart, and Senora formations.[19] The individual cyclic successions in some instances are several hundred feet thick.

Conditions in north-central Texas are

[17] John M. Jewett, "Classification of the Marmaton Group, Pennsylvanian, in Kansas," *Univ. Kan. Pub., Kan. State Geol. Surv. Bull. 38*, Part XI (1941), pp. 285–344.

[18] R. C. Moore, "Stratigraphic Classification of the Pennsylvanian Rocks of Kansas," *Univ. Kan. Bull. 22* (1935), p. 32.

[19] C. H. Dane, H. E. Rothrock, and J. S. Williams, "Geology and Fuel Resources of the Southern Part of Oklahoma Coal Field. III. The Quinton-Scipio District," *U.S. Geol. Surv. Bull. 874-C* (1938), pp. 157–68.

what like those in southern Oklahoma, but the succession is thinner, and coal is less common.[20] The example shown in column 7 (Fig. 3) includes the Palo Pinto limestone and underlying shale and sandstone at the top of the Strawn and the base of the Canyon formations of Missouri age. As in northeastern Oklahoma, the limestones, where sufficiently thick, tend to cap eastward-facing escarpments. The disappearance of coal in this and most columns farther west was perhaps due to somewhat less humidity than in the eastern states, where coal is more prominent.

Column 8 (Fig. 3) represents beds of Gaptank age in the Marathon Basin, western Texas.[21] It is somewhat similar to column 7 from north-central Texas, but the limestone is thicker; there are no argillaceous beds, or only a slight thickness of them; and some of the sandstones are conglomeratic, containing slightly weathered, poorly rounded fragments, which were evidently locally derived. This type of sediment, which is also widespread in the southern Rocky Mountains, results from the fact that the depositional basins were nearer land areas that possessed some relief than were the basins farther east.

Column 9 (Fig. 3) represents a type of cycle found in the Big Blue series, Lower Permian, of Kansas and Nebraska.[22] It consists wholly of marine shale and limestone but shows a definite cyclical pattern in the lithology and fauna of the shales and limestones. It may be contrasted with Figure 1, column 9, and Figure 2, column 7, both of which are wholly nonmarine.

TYPES OF PENNSYLVANIAN CYCLES IN THE ROCKY MOUNTAINS

In the Rocky Mountains the Pennsylvanian lacks coal, carbonaceous shale, underclay, and "fresh-water" limestone almost entirely, and shale and ironstone are uncommon (Fig. 4). Although the lithology at most places exhibits a pattern suggestive of some sort of rhythmic deposition, detailed studies by C. B. Read and others in New Mexico show that the cyclic units are much less widespread than in the central and eastern United States, and thin members have not yet been widely traced and correlated anywhere in the Rocky Mountains. Although this greater variability may result from localized lands with high relief adjoining local depositional basins in parts of the Rocky Mountains, elsewhere it probably reflects less detailed stratigraphic studies than have been carried on farther east; and the numerous fossiliferous zones in such areas as the Hartville Uplift, the Black Hills, the Laramie Range, central and western Colorado, and the San Juan Mountains of southwestern Colorado would seem to offer the opportunity for precise correlation of thin members through large areas after adequate sections have been carefully studied lithologically and faunally.

In Figure 4, column 1 represents a type of alternation in the "Brazer" limestone, later or Middle Mississippian of western Wyoming. The sequence displays dense, dark, lithographic, nearly unfossiliferous limestone, overlain by lighter or pinker limestone breccia with angular or subangular limestone pebbles. Several such alternations seem to be per-

[20] B. Plummer and Joseph Hornberger, Jr., "Geology of Palo Pinto County, Texas," *Univ. Tex. Bull. 3534* (1935), pp. 31–32.

[21] Philip B. King, "The Geology of the Glass Mountains, Texas. I. Descriptive Geology," *Univ. Tex. Bull. 3038* (1930), p. 44.

[22] John M. Jewett, "Evidence of Cyclic Sedimentation in Kansas during the Permian Period," *Trans. Kan. Acad. Sci.*, Vol. XXXVI (1933), pp. 137–40.

sistent in the Gros Venture, Teton, Hoback, Wyoming, and Snake River ranges of western Wyoming and eastern Idaho. Their significance is not known. It has been suggested[23] that the breccias may be founder breccias resulting from the solution of interbedded gypsum or anhydrite, but no field evidence in support of this theory has been discovered.

Column 2 (Fig. 4), characteristic of the Tensleep and Quadrant formations of northwestern Wyoming and adjacent parts of Montana, consists of a two-member alternation of dense, hard, well-sorted quartzitic sandstone and light-gray to white, dense, somewhat cherty, dolomitic limestone. Although this type of alternation appears in several mountain ranges, the individual pairs of beds have not yet been traced between ranges, so that it is not known whether the cycle records local or widespread fluctuations of some sort. Minor erosional unconformities with a relief of a few inches are common at the bases of the sandstones. Many of the limestones and dolomites are wholly unfossiliferous, and fossils are rare and fragmentary throughout the section.

Columns 3, 4, and 5 (Fig. 4) show, respectively, representative alternations in the Minnelusa formation, Black Hills, South Dakota; the Guernsey formation, Hartville Uplift, Wyoming; and the Casper formation, Laramie Range, Wyoming. The Black Hills and Laramie Range sections show alternations of marine fossiliferous limestones with sandstones, the sandstones in the Laramie Range being conglomeratic in some cycles. In the Hartville area the alternation is between limestone and red calcareous shale.

Column 6 (Fig. 4) is representative of the Fountain formation of the Denver region and was measured at Red Rocks Park, Morrison, Colorado. Here the Fountain formation, which seems to be wholly nonmarine and barren of all fossils, displays an alternation of pink arkosic conglomerate and maroon shale, with evidence of slight disconformity at the top of each shale. There is no way to determine whether these alternations correspond approximately with cyclothems of the eastern states, nor is it known whether the pairs of beds achieve wide areal extent within the Denver–Colorado Springs region.

Columns 7 and 8 (Fig. 4) are in the Salida region, southern Colorado, on the east side of the Sangre de Cristo Range. In that region the lowest few hundred feet of the Weber formation, equivalent to the Glen Eyrie beds of the Colorado Springs area, contain quartzose sandstone, underclay, thin coal, and plant-bearing nonmarine shales (col. 8), the whole succession resembling that in the Lower Pennsylvanian of western Illinois. Similar beds occur in the Lower Pennsylvanian near Pecos, New Mexico. These are the only Rocky Mountain Pennsylvanian beds displaying carbonaceous sediments like those in the eastern states. Above these beds the succession changes abruptly to a type of cycle repeated many times in the Upper Weber and Maroon formations (col. 7). This succession includes arkosic conglomerate, sandstone, and marine limestone, with comparatively little clay shale and no coal or underclay. This succession resembles that of other parts of the southern Rocky Mountains (cols. 9, 11, and 12), but with variations in the coarseness of conglomerate or sandstone. The change in character between columns 7 and 8 at the same locality may be attributed to a decrease in humidity, so that the rocks in the upland

[23] M. Y. Williams, personal communication.

areas supplying sediment to this basin were weathered more physically than chemically.

Column 9 (Fig. 4) is in the Sangre de Cristo formation at La Veta Pass, Colorado, farther south in the Sangre de Cristo Range. Some of the conglomerates of this region are very coarse, the individual cobbles reaching a maximum size of about 6 feet. Many of the fragments are granite or fresh feldspar.

Column 10 (Fig. 4) represents the Molas formation, Lower Pennsylvanian, in the San Juan Mountains, southwestern Colorado. This formation, although thin on the outcrop, shows several alternations of thin, nonarkosic, round-pebble conglomerates with red shales, in what seems to represent some sort of cyclic deposition. The succession is unfossiliferous and resembles that of the Fountain formation near Denver (col. 6) except that the Fountain formation is arkosic and the pebbles are generally more angular than those in the Molas conglomerates.

Column 11 (Fig. 4) is representative of the Hermosa formation, overlying the Molas in the San Juan Mountains. It is somewhat similar to the Sangre de Cristo section at La Veta Pass (col. 9) except that the sandstones are finer grained and generally nonconglomeratic.

Column 12 (Fig. 4) is from the southern part of the Sangre de Cristo Range near Taos, New Mexico, where the strata are referred to the Magdalena formation. Although the individual beds of the Taos succession are reported to be quite lenticular, there appear to be at least eighty cycles, averaging 60–80 feet thick. Like the La Veta Pass section (col. 9), some of the conglomerates include cobbles as large as 6 feet in dimension. There are, however, more numerous and more prominent marine fossiliferous limestones than in the La Veta Pass area.

Column 13 (Fig. 4) is from the Magdalena formation near Pecos, New Mexico. The succession differs from that at Taos (col. 12) in that the sandstones and conglomerates are thinner and finer grained and the limestones are thicker. The limestones here produce prominent benches and escarpments, as they do in parts of Texas and Oklahoma.

Column 14 (Fig. 4) represents a portion of the Paradox formation near Moab, eastern Utah, between the Molas (below) and the Hermosa formations. The column is based on a partial section taken from a diamond-drill core.[24] The core record seems to indicate a fairly regular cyclic repetition of shale, limestone, gypsum, and salt through a total thickness of about 6,000 feet. This is the only cycle that includes important evaporite beds and may not in any way correspond to the other cyclic successions included in Figure 4. The salt is interpreted as the final unit of the cycle, followed by shale with an increase in the amount of sediment washed into the barred basin in which the Paradox formation was deposited.

There are, no doubt, other areas exhibiting still other types of cyclic deposition, such as the cycle described from the Permian Kaibab limestone of the Grand Canyon region.[25]

Regions studied by the author that do not seem to display any form of cyclic sedimentation are southeastern Idaho, where the Wells formation is nearly solid limestone, and the Wasatch Moun-

[24] A. A. Baker, "Geology and Oil Possibilities of the Moab District, Grand and San Juan Counties, Utah," *U.S. Geol. Surv. Bull. 841* (1933), p. 86.

[25] Edwin D. McKee, "The Environment and History of the Toroweap and Kaibab Formations of Northern Arizona and Southern Utah," *Carnegie Inst. Wash. Pub. No. 492* (1938), pp. 129–32.

tains of Utah, where the Weber formation is mostly uniform fine-grained quartzitic sandstone. The Naco formation of southeastern Arizona may or may not be cyclic but is not classified here because it has not been seen by the author. It is mostly limestone. Little information is at hand as to the cyclic or noncyclic character of Pennsylvanian sediments in the Great Basin, Canadian Rockies, and Pacific Coast areas.

PERSISTENCE OF PENNSYLVANIAN BEDS

The remarks thus far have stressed the vertical and lateral heterogeneity of the Pennsylvanian system. It should be noted in contrast that some very thin beds attain almost unbelievable lateral extent. For example, the fire-clay coal of eastern Kentucky and its correlatives in other states have a 4–6-inch parting of flint fire clay, which has been traced through considerable portions of the West Virginia, Virginia, Kentucky, and Tennessee fields.[26] Even more extensive is the parting known as the "Blue Band" in the Herrin (No. 6) coal of Illinois. This and one or two other thinner partings are found wherever the coal is known and at some places in Illinois, Indiana, western Kentucky, Iowa, and Missouri, where the coal itself is represented only by smut streaks; and available evidence suggests that the coal may be equivalent to the Middle Kittanning of Ohio, Pennsylvania, eastern Kentucky, and West Virginia, which at most places displays a similar succession of thin partings. The average thickness of the individual partings in this coal is less than 2 inches, and in some places the thinnest is plainly recognizable with a thickness less than $\frac{1}{4}$ inch.

A 2–6-inch band of cone-in-cone attached to the principal limestone bed of the Oak Grove member, Liverpool cyclothem, has been traced from the Rock Island region of northwestern Illinois to the Terre Haute region of western Indiana. The Covel conglomerate,[27] a bed of limestone conglomerate 1–6 inches thick between coals Nos. 4 and 5, has been found at many localities in Iowa, Illinois, and Indiana. Many other examples of the great persistence of thin beds could be cited, especially from the eastern Interior and mid-continent areas.

GRAPHIC REPRESENTATIONS OF VARIATIONS IN PENNSYLVANIAN LITHOLOGY

It seems that some suitable method might be devised to show graphically the nature of lateral change in lithology of the Pennsylvanian and its subdivisions. In approaching this problem the author prepared a sheet (Fig. 5) to record data pertaining to the lithology of Pennsylvanian rocks of a limited area. A preliminary count showed that data pertaining to the Pennsylvanian might be available for about eleven hundred counties in which the system either outcrops or lies at a sufficiently moderate depth to be readily accessible to the drill. It was therefore decided to use the county as the unit of variation except where a county is divided into segments by a major structural feature like the La Salle anticline of Illinois or the Pine Mountain thrust fault, in which cases two or more sheets are prepared for a county. The sheet provides a place to record the locality, source of data and reliability of the column; the age of overlying strata, if any, and strata immediately below the

[26] H. R. Wanless, "Pennsylvanian Correlations in the Eastern Interior and Appalachian Coal Fields," *Geol. Soc. Amer., Spec. Paper 17* (1939), pp. 55–57.

[27] H. B. Willman, "The Covel Conglomerate, a Guide Bed in the Pennsylvanian of Northern Illinois," *Trans. Ill. Acad. Sci.*, Vol. XXXII (1939), pp. 174–76.

DATA SHEET, PENNSYLVANIAN LITHOLOGY

State: W. VIRGINIA *County:* KANAWHA *Locality of column:* COMPOSITE—FROM SIX COLUMNS
Reliability: GOOD *Reference:* *W.Va. Geol. Surv. County Repts., Kanawha County*
Overlying stratum: DUNKARD PERMIAN *Substratum:* MAUCHCHUNK MISSISSIPPIAN

Rock Type	Virgil			Missouri			Des Moines			Lampasas			Morrow			Stanley Jackfork			Total		
	N.	T.	%	N.	T.	%	N.	T.	%	N.	T.	%	N.	T.	%	N.	T.	%	N.	T.	%
Arkosic cgl.																					
Quartz cgl.										2	4′	0.7	6	72′	7.9	1	1′	0.2	9	76′	2.4
Sandstone	8	228′	43.2	6	225′	65.2	8	435′	89.1	14	271′	51.5	19	607′	67.5	4	155′	42.6	59	1,921′	61.4
Red ss.																					
Oil zone																					
Plant sh.	11	166′	32.9	3	62′	17.9	2	20′	4.2	16	128′	24.6	11	166′	18.4	4	111′	30.5	47	654′	21.
Marine sh.										3	25′	4.7							3	25′	0.8
Black slate																					
Dark shale	1	1′	0.1							6	53′	10.1	7	53′	5.9	5	97′	26.4	19	203′	6.5
Red shale	9	102′	21.9	4	48′	13.9													13	150′	4.8
Fire clay							1	6′	1.3										1	6′	0.2
Underclay	2	5′	1.1	3	6′	1.7	2	3′	0.6	5	5′	1.0				1	3″	0.1	19	19′	0.6
F.W. ls.																					
Marine ls.				1	2′	0.6				4	4′	0.8							5	6′	0.2
Fusulinid																					
Dolomite																					
Cherty ls.																					
Chert							1	4′	0.8										1	4′	0.1
Coal	1	5′	0.9	2	2′	0.7	7	18′	3.7	14	35′	6.6	3	3′	0.3	2	7″	0.2	29	63′	2.
Iron ore																					
Gypsum																					
Salt																					
Lava																					
Pyroclastic																					
Schist																					
Gneiss																					
Total	32	507′	100	19	346′	100	21	46′	100	64	524′	100	46	901′	100	17	364′	100	199	3,128′	100
Cycles	11			5			5			13			5			5			44		

V — M — D — L — M — SJ

Monongahela | Conemaugh | Allegheny | Kanawha | New River | Pocahontas | Princeton

FIG. 5.—Data sheet for Kanawha County, West Virginia, recording proportions of various rock types in the Pennsylvanian column for this country

Pennsylvanian; the part of the total Pennsylvanian column represented in the county in terms of the major subdivisions used in the National Research Council Committee report—Virgil, Missouri, Des Moines, Lampasas, and Morrow—with the addition of a basal Pennsylvanian or Pennsylvanian-Mississippian unit, here called the "Stanley-Jackfork," the names locally applied to principal subdivisions; and the number of beds (*N*), total thickness (*T*), and percentage (%) of 26 kinds of rocks in each of the divisions, as well as in the total Pennsylvanian section for the county. The beds comprising each of these divisions are determined in accordance with the correlation chart accompanying the report of the Pennsylvanian subcommittee.[28] If the section is cyclic, the number of cyclic units in each subdivision of the Pennsylvanian is recorded. With intensive application for about 1 month and with student assistance,[29] made available by the University of Illinois graduate school, data sheets like that in Figure 5 have been prepared for about one hundred and sixty counties, including all the Pennsylvanian areas of Georgia, Tennessee, Virginia, West Virginia, and Maryland and parts of eastern Kentucky, Ohio, Illinois, and Missouri. Further studies of this type are contemplated, directed toward the completion of data sheets for all Pennsylvanian areas in the United States. Assistance will be greatly appreciated in the form of access to detailed descriptions of the Pennsylvanian of any area from outcrops, core descriptions, or sample studies. Such sources have been used almost exclusively except in northwestern West Virginia and parts of eastern Kentucky, where the best available information on part of the subsurface Pennsylvanian was in the form of drillers' logs of oil wells. Some difficulty has been experienced in classifying shales as plant (nonmarine), marine, or dark. Some records and descriptions fail to mention underclays and thin coals, while others undoubtedly refer to some common shales as fire clays. The pebbly character of thick massive sandstones is generally better indicated in outcrop descriptions than in subsurface records. The sheet displayed (Fig. 5) represents Kanawha County, West Virginia, the only county thus far encountered in which the entire Pennsylvanian seems to be recorded. The range in length of column is from 10 feet in Whiteside County, northwestern Illinois, where a sandstone of Des Moines age fills a channel cut in Silurian dolomite, to Wise County, Virginia, where there are 7,088 feet extending up from the base of the Pennsylvanian only about as high as the 10-foot sandstone of Whiteside County. In Wise County, where excellent outcrops and core-drill records are the source of information, there are 88 coal beds, aggregating 132 feet 6 inches and totaling 1.8 per cent of the Pennsylvanian column; 112 sandstones, aggregating 2,412 feet; and 152 nonmarine shales, totaling 3,396 feet. In contrast, there is only one marine limestone, 8 inches thick; and there are 6 marine shales (fossil-bearing), totaling 66 feet.

The figures representing numbers of beds, total thickness, and percentage can be plotted and contoured as lithologic-variation maps, and the total thickness of each division of the Pennsylvanian can be used to prepare isopach maps.

Since most of the available time has

[28] R. C. Moore and others, "Correlation of Pennsylvanian Formations of North America," *Bull. Geol. Soc. Amer.*, Vol. LV (1944), chart opp. p. 706.

[29] The service rendered by Miss Lorena Ross and Mr. Robert Knodle, research assistants, and Miss Norma Wallis, draftsman, aided greatly in compiling the data and preparing the illustrations accompanying this paper.

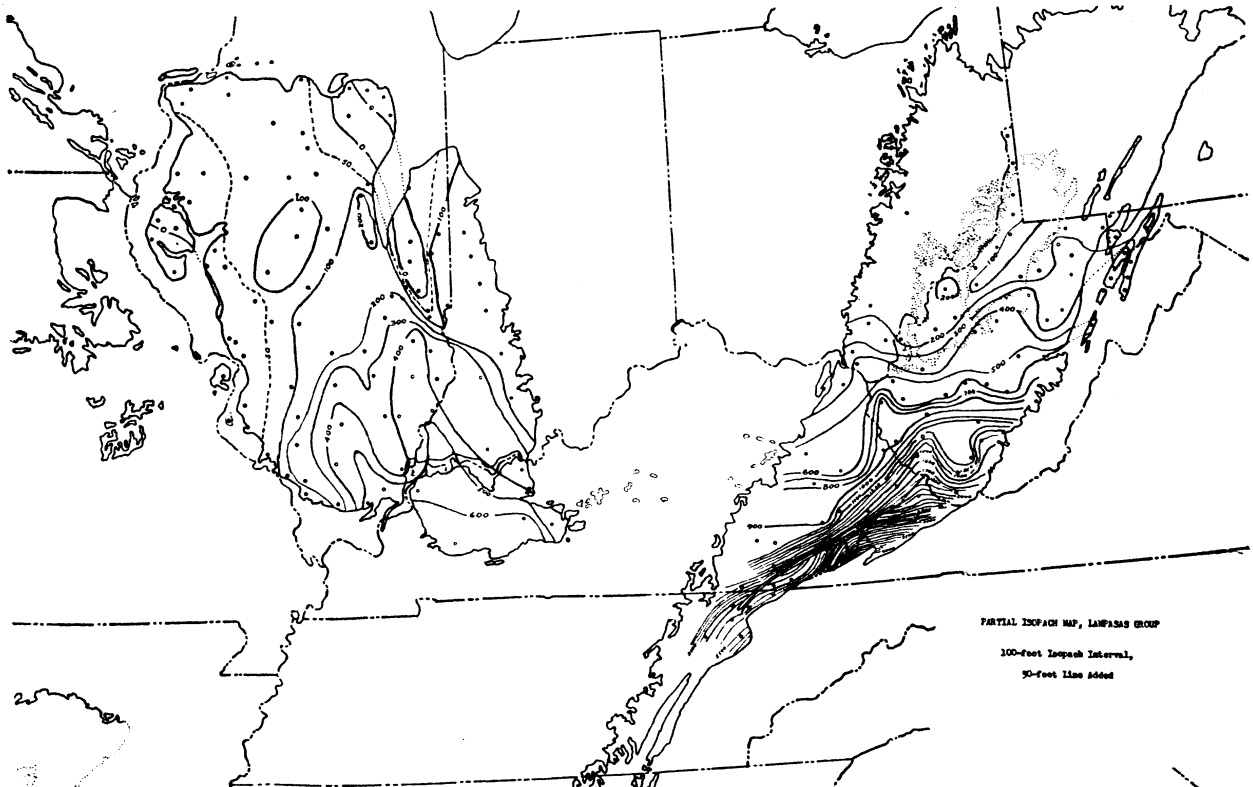

FIG. 6.—Partial isopach map of the Lampasas Group for the eastern Interior Basin and part of the Appalachian Basin. Isopach interval 100 feet.

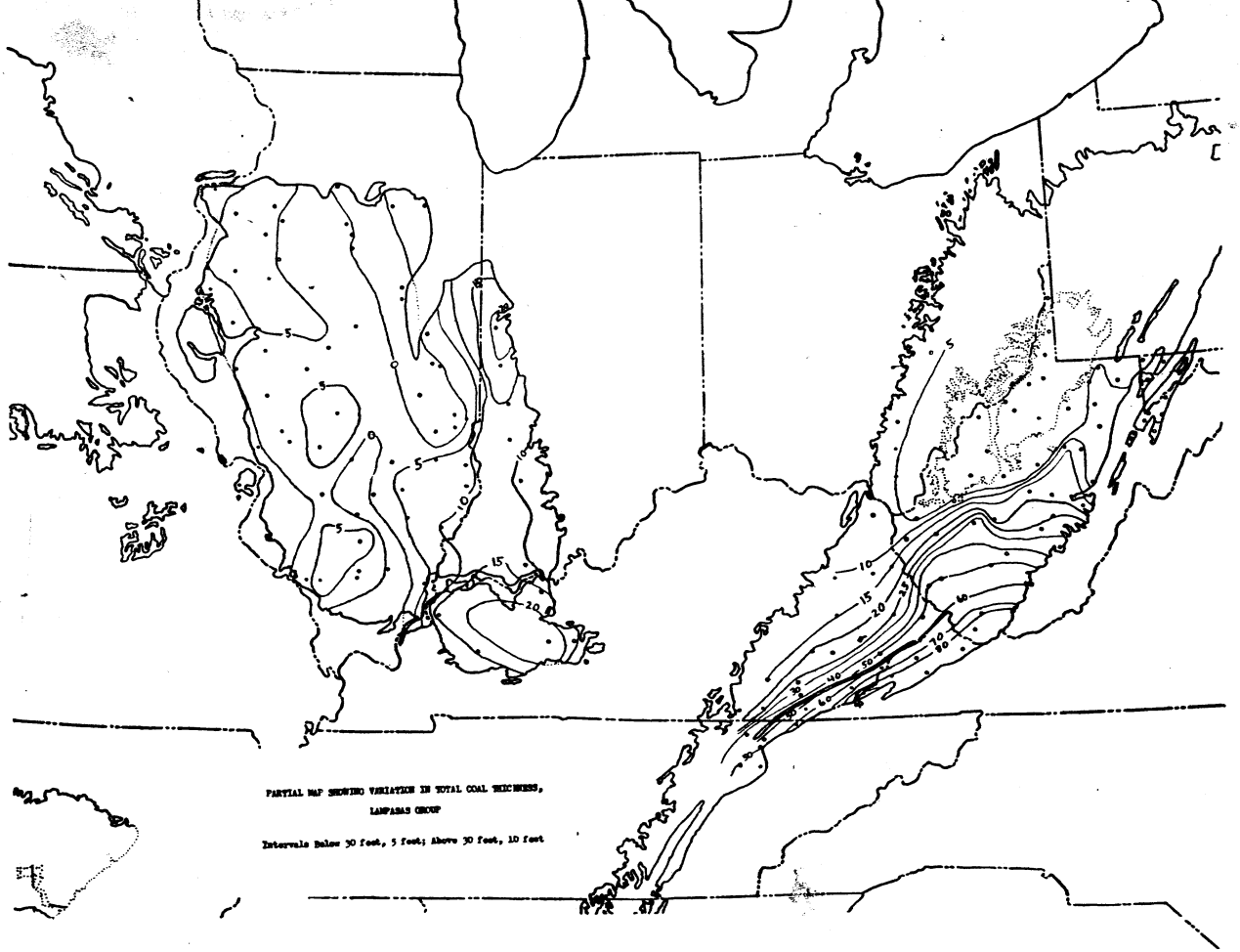

FIG. 7.—Partial map showing variation in total coal thickness, Lampasas group. Intervals below 30 feet, 5 feet; above 30 feet, 10 feet

been devoted to the assembly of lithologic data, only two partial maps have been prepared. The first (Fig. 6) is an isopach map of the Lampasas group in parts of the Appalachian and eastern Interior basins, the boundaries used in the several states being those in the correlation chart prepared by the Pennsylvanian subcommittee of the Stratigraphic Committee of the National Research Council. This map shows a variation from zero in certain areas in Illinois, in which the earliest Pennsylvanian strata are post-Lampasas, to 4,080 feet in Lee County, Virginia, near Pennington Gap. The two areas of zero thickness are (1) along the La Salle and Oakland anticlines of eastern Illinois and (2) an area in extreme western Illinois from near Quincy southeastward toward St. Louis. In each of these areas positive tendencies, perhaps associated with post-Mississippian movements, caused the area to be sufficiently high to prevent Pennsylvanian deposition until after the Lampasas. In the eastern Interior Basin a zone of notable, though moderate, thickness, 200–300 feet, is the trough just west of the La Salle anticline. In the Appalachian field the thickness ranges from about 150 feet in southern Ohio to a maximum of 4,000 feet near the southeastern border of the field. In the southeasternmost counties this group expands southeastward at the rate of at least 100 feet per mile. The abrupt thickening on the south, as compared with the north side of the Pine Mountain thrust fault, suggests that this structure developed along a hinge line of differential downwarping which existed during Lampasas time. When data sheets have been prepared for the Ozark-Ouachita area in Arkansas and Oklahoma, it will be possible to compare the pattern of expansion there with that in Kentucky, West Virginia, and Alabama.

The only map of lithologic variation thus far prepared (Fig. 7) shows the total thickness of Lampasas coals. The coals range from zero thickness to a maximum of 91 feet, with 37 coals in Lee County, Virginia, and an aggregate of 87 feet in adjacent Wise County. In the Appalachian field the pattern of total coal-thickness variation is similar to that for the Lampasas total thickness as shown on the isopach map (Fig. 6), but there is an area in northern West Virginia with no coal, or at least less than 5 feet of coal. Although most of the source records in the several counties reporting no Lampasas coal are drillers' logs, a diamond-drill boring near Wheeling reports only 2 feet of coal and indicates that there is really very little Lampasas coal in that area. In the eastern Interior Basin the maximum coal, about 20 feet, is in the western Kentucky field, with a secondary high in the northern part of the Indiana Pennsylvanian area, where, in the Brazil Block coal field, the Minshall and Upper and Lower Block coals of Lampasas age are all minable. A map showing variation in the percentage of coal in the Lampasas group has been undertaken but is not yet completed. Another type of map, which might be useful and will be tried, is a map with a circle divided into segments to show percentages of sandstone, shale, and limestone or some other ratios of abundance of different rocks.

It is believed that numerous broadly regional relationships between variations in thickness, lithologic variation, and tectonic features will be evident as a result of further graphic studies like those suggested here. These may be considered simply preliminary samples of what may be accomplished. Lithologic-variation studies should contribute to our understanding of Pennsylvanian paleogeography, climate, and tectonic history.

INVERTEBRATES IN PENNSYLVANIAN CORRELATIONS

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ABSTRACT

Comparatively few Pennsylvanian fossils are known to have precise zonal significance, and consequently most species at present have no more than general correlative value. This is partly because the Pennsylvanian stratigraphic sequence is uninterrupted, except locally, by important breaks separating easily distinguished faunas and partly because most Pennsylvanian species are too poorly understood to be recognized with precision and separated from closely related forms. Complications are introduced by the extreme diversity of Pennsylvanian environments, which resulted in the close stratigraphic association of almost wholly different faunules.

The faunal approach to Pennsylvanian correlation has produced results that are unsatisfactory except in a most general way. Detailed correlations can be made much more accurately on the basis of lithologic sequence, and studies of this type have revealed notable errors in correlations based on fossils. Moreover, faunal studies have resulted in the misidentification of certain diagnostic species and confusion regarding the ranges of others.

The detailed study of closely related fossils from many different stratigraphic zones in an effort to work out evolutionary trends gives great promise of much more useful and significant results. The establishment of sequences of evolutionary stages is now known to be possible and offers a means of correlation far superior to that which is dependent on the mere presence or absence of individual index species. All such studies are still in their preliminary stages, but more or less progress has been made with several groups, particularly the fusulinids.

INTRODUCTION

Some thousands of invertebrate species have been described from the Pennsylvanian rocks of America. A very few of them have precisely known stratigraphic significance, a somewhat larger number is believed to have more or less general correlative value, but the great majority is at present of little use to the careful stratigrapher.

This situation is by no means unique, although the ranges of fossils in several other systems are more completely known. Like those of other ages, many Pennsylvanian species were originally described from specimens whose exact localities and stratigraphic positions are now unknown. Old concepts regarding species were elastic, and original descriptions are generally inadequate for modern precise discrimination. These and other deficiencies are problems that have been inherited by present-day stratigraphic paleontologists.

Most other Paleozoic systems have particularly favored general localities where the succession of faunas has been worked out in more or less detail and where a considerable number of guide fossils has come to be recognized. At many of these places the stratigraphic section has obviously been interrupted, or important changes in environment have occurred, so that there are clearly marked and extensive breaks between distinctive faunas.

This is not true of the Pennsylvanian. Fossil invertebrates are rare in the type region of this system, and the intricacy of its stratigraphic details has discouraged careful zonal investigations in other areas until quite recently. Although the stratigraphic limits of a few common Pennsylvanian species have been determined, no faunal breaks comparable to those in some of the other systems have been recognized, and, with a very few more or less local exceptions, it is now

fairly certain that no such breaks exist. On the other hand, environments varied so greatly both from place to place and from time to time that almost entirely distinct faunules occur laterally in a single zone or in two adjacent beds.

At first glance this situation seems most discouraging, and undoubtedly numerous stratigraphic paleontologists, as a result, have directed their attention to other and apparently more promising fields. However, recent developments in Pennsylvanian stratigraphy, which make possible the accurate recognition and identification of many thin zones throughout wide areas, have completely changed this picture. It is now possible to assemble many collections from widely scattered localities, all precisely dated with respect to one another, in a way that cannot be duplicated in any other system. This means that assemblages of fossils can be studied and directly compared from numerous distinct successive zones arranged at reasonably short intervals throughout a very considerable thickness of strata and spanning a very considerable length of time, unbroken by any important hiatus. Thus, if evolution were progressive in any lineage, it should be possible to follow it step by step and work out certain stages indicative of definite stratigraphic zones. The recognition of such evolutionary stages would find direct application in correlations more secure than those based on the mere presence of some species and the absence of others.

PROBLEMS INVOLVED

The problems facing Pennsylvanian stratigraphic paleontologists are similar to those encountered in the study of fossils from other systems. They are of two main types: first, the establishment of already named species and, second, the recognition of constant variants, some of which may be expected to fall within well-marked evolutionary series.

It is assumed, of course, that adequate suites of specimens are available for study. These may be old collections or new, but it is absolutely essential that they be accurately located in the stratigraphic section. Exact stratigraphic position may be difficult or impossible to establish or may be more or less doubtful for old collections; hence new ones are preferable in every way. They must be carefully made in strict accordance with the most modern practices of Pennsylvanian stratigraphy, and particular care must be taken to avoid mixing of specimens from adjacent zones. Old collections of somewhat doubtful origin or new ones that for one reason or another cannot be accurately located stratigraphically are not useless, however, because they can be employed to check observations and conclusions made on other material.

FAUNAL STUDIES

The study of more or less local but complete faunas has been the method of approach to correlation by many paleontologists during the last fifty or more years, and the degree of time equivalence of different beds or formations was, and still is, believed to be indicated by the percentage of common species present in any two related faunas. This percentage method of correlation has yielded many significant and valuable results; but it is quite obvious that conditions of environment introduce complicating factors that have been considered far too little. This method may work satisfactorily in a stratigraphic province where uniform conditions were widespread and persisted for a considerable length of time or in areas where the stratigraphic section is interrupted by important breaks. Sev-

eral of the older Paleozoic systems in east-central United States meet these conditions to a greater or less extent, and correlations of their subdivisions have been made in this way with reasonable success.

The percentage method of correlation, however, fails completely in the Pennsylvanian system because of the lack of uniformity in environment. It is common knowledge that the faunules of two beds of like lithology, separated by many feet of intervening strata, are likely to contain nearly identical species, while two dissimilar adjacent beds may have almost no species in common.

Five different general types of Pennsylvanian marine faunules may be distinguished. These are: (1) characteristic limestone faunules in which molluscoidean species greatly predominate, the type of faunule with which fusulinids are commonly associated; (2) algal-limestone faunules in which mollusks predominate; (3) molluscan shale faunules; (4) black-shale faunules, also largely mulluscan, but quite different from the last. No single species occurs commonly in all five types of faunules, and many species appear to be strictly limited to faunules of one particular type.

The above rough classification is certainly arbitrary, and many intergradations occur between the faunules listed. They are all generally related to certain lithologically different, but distinctive, types of strata, and the conclusion that they were controlled by different combinations of environmental conditions can hardly be avoided. Also there are other types of faunules. Some appear abnormal because the presumably controlling factors of environment are not plainly indicated by lithology. For example, differences in salinity strongly influence the composition of a faunule, without effecting any comparable change in sedimentation.

Obviously, correlations made by the percentage method and based upon such diverse faunules can be only a source of confusion and error.

Another shortcoming of paleontological studies by faunas concerns the accurate determination of species. In his investigation of a fauna a paleontologist first sorts his specimens into groups that he considers to be species. Then he compares each group, in turn, with specimens already identified or with descriptions and illustrations in the literature. He notes the similarities and differences, and for each group he must decide whether his specimens are sufficiently similar to described forms to be referred to the same species or different enough to be considered new species or varieties. Such decisions require the exercise of sound judgment, and the conclusions reached by an experienced investigator are much more likely to stand the test of time than are those of a student without broad acquaintance with related faunas. When approached in this way, however, the conclusions of the most experienced paleontologist are open to more or less serious question on two different counts. In the first place, because he is unacquainted with the type specimens, he may be misled by imperfect or inadequate descriptions and illustrations and misidentify his specimens. Second, he is rarely able to evaluate the significance of observed minor differences and cannot be sure whether they are individual variations within a species, variations induced in response to local environmental factors, or variations that mark stages in an evolutionary series. Moreover, his misidentifications or errors in judgment are likely to be perpetuated indefinitely by later workers, even after more careful or

more complete studies have revealed their existence.

G. H. Girty's report[1] on the Wewoka fauna of Oklahoma is excellent and is probably the most useful single publication on American Pennsylvanian invertebrates. Because it is a faunal study, however, it exemplifies some of the shortcoming and faults mentioned above. For example, his specimens of *Chonetes mesolobus* do not correspond with the original description of that species based on specimens collected near St. Louis, and, consequently, he referred them to two new varieties. To indicate the differences, he illustrated and discussed specimens from Ohio that possess the characters mentioned in the original description. His conclusions were amply justified on the basis of the information available to him and to everyone else at that time. Unfortunately, however, the original description is in serious error. His specimens are comparable to the originals, and the Ohio specimens represent a distinctly different species characteristic of a different paleontologic zone in the Lower Pennsylvania.[2]

Also, the Wewoka specimens referred by Girty to *Phanerotrema grayvillensis* are not exactly similar to those occurring at the type locality, which is in a zone much higher stratigraphically. The differences are not striking, but they are constant and, once noted, easily recognized. Specimens from intermediate zones indicate that the originals and the Wewoka forms represent different stages in a single evolutionary line. Girty noted certain differences, but he was ignorant of their significance and concluded that they are unimportant. Actually, his specimens of so-called *Phanerotrema* are much more worthy of a distinctive name than either of his varieties of *C. mesolobus* because the former are characteristic of different stratigraphic zones and are useful for correlation, whereas his varieties of the latter are not.

GUIDE FOSSILS

A good guide fossil should be a common species, otherwise its absence from a faunule may have little significance. It should have been able to live under a variety of environmental conditions, so that its occurrence is not closely restricted ecologically. Also it should have short stratigraphic range. Unfortunately, there are few Pennsylvanian fossils that meet these requirements. Almost without exception, the recognized species that are common and likely to be present in various different kinds of rock were such successful and adaptable forms that they persisted for considerable lengths of time and consequently are not useful for close correlation.

Some species of this type characterize definite parts of the Pennsylvanian succession, like *Mesolobus* and several brachiopods commonly referred to *Marginifera*, which are very abundant in certain zones but also occur in smaller numbers in both higher and lower beds and whose exact stratigraphic ranges have not been accurately determined. Nevertheless, they are characteristic of several zones, such as those distinguished as series on the Pennsylvanian correlation charts;[3] but such zones are rarely sharply delimited from one another and commonly overlap. Because of these relations, however, more accurate correla-

[1] "Fauna of the Wewoka Formation of Oklahoma," *U.S. Geol. Surv. Bull. 544* (1915).

[2] J. M. Weller and R. McGehee, "Typical Form and Range of *Mesolobus mesolobus*," *Jour. Paleon.*, Vol. VII (1933), pp. 109–10.

[3] R. C. Moore and others, "Correlation of Pennsylvanian Formations of North America," *Bull. Geol. Soc. Amer.*, Vol. LV (1944), pp. 657–706.

tions can be made on the basis of the associations of certain common species than by noting the occurrence of any single species.

In recent years the fusulines have been studied more intensively than any other group of Pennsylvanian fossils. A considerable number of closely related genera of wide distribution and restricted stratigraphic range have been recognized by C. O. Dunbar, M. L. Thompson, and others. Although their exact ranges are still more or less incompletely known, they furnish the best means of interbasin correlation now generally available. Even when species are not identical, significant conclusions may be reached by comparing stages in the progressive development of certain structures.

Other microfossils are as yet of little value for Pennsylvanian correlations. Recent work on the ostracodes by C. L. Cooper[4] gives promise of greater usefulness for these fossils; but the smaller foraminifera and conodonts have not been sufficiently investigated for their potentialities to be known.

Sponges may be useful for local correlations in certain areas, but they are neither common enough nor widely enough distributed to be of any general importance.

The study of Pennsylvanian corals has been greatly neglected, and most specimens are commonly identified as one or another of a few old species, all of which require restudy. The work of R. M. Jeffords and R. C. Moore on Permian and Pennsylvanian corals and of W. H. Easton on Mississippian corals has shown the possibilities of modern investigations by sectioning methods; and Pennsylvanian forms offer a research field that is almost certain to produce important results.

Crinoids, although interesting and undoubtedly significant, are too rarely represented by well-preserved specimens to be of particular importance for correlation. The investigations of R. C. Moore, however, are adding greatly to knowledge of these fossils.

Bryozoans, like corals, have long been neglected. They are, however, very abundant in the Pennsylvanian at many places and exhibit a diversity of form and structure unsurpassed by any other fossils. They are now being studied by G. E. Condra and M. K. Elias and offer great promise as useful guide fossils when they are better known.

The brachiopods include more standard index fossils of the Paleozoic than any other phylum, and, until recently, brachiopods were largely relied upon for correlations in the Pennsylvanian. They are still more generally useful than any other group of fossils.

The variety of brachiopods present in the Pennsylvanian system is well indicated by the work of C. O. Dunbar and G. E. Condra.[5] They list 133 species and varieties referred to 42 genera; but they consider Lower Pennsylvanian forms incompletely. These figures are a good indication of the decline of the brachiopods after the Mississippian, in which system they are represented by more than twice as many genera and species.

Although numerous species of Pennsylvanian brachiopods are locally abundant, few of them are of general common occurrence, and most of these possess more or less elastic specific limits and range through considerable, but not sharply delimited, thicknesses of strata. Consequently, brachiopods are not at

[4] "Pennsylvanian Ostracodes of Illinois," *Ill. Geol. Surv. Bull.* 70 (1946).

[5] "Brachiopoda of the Pennsylvanian System in Nebraska," *Neb. Geol. Surv. Bull.* 5 (2d ser., 1932).

present particularly useful for close correlations, although they do serve admirably to mark certain general stratigraphic zones.

Dunbar and Condra's report does not indicate that they attempted to work out lines of evolutionary development in the brachiopods. Because of the very common occurrence of some forms, however, such an attempt would certainly be worth while. Unfortunately, brachiopods lack great variety in surface ornamentation and possess few other external characters, other than general shape and form, that, in combination, might distinguish significant and more or less constant varieties. The two groups that seem to offer the most promise are the chonetids and some forms generally referred to *Marginifera*.

Paleozoic pelecypods present numerous difficulties to their investigators. These arise primarily because such important features as hinge structures are rarely seen and because many forms resemble one another in shape and lack distinctive surface markings. Furthermore, the shape of certain forms varied considerably at different stages of growth. In proportion to the number of species that have been described, pelecypods are probably the most inadequately known Pennsylvanian fossils.

Pelecypods are locally very abundant in Pennsylvanian strata, but mostly they appear to have been more than ordinarily circumscribed in their habitat relations. At present they are of negligible value for correlation, but the studies of N. D. Newell indicate that detailed investigations may be expected to produce useful results in this respect.

Gastropods also were, for the most part, closely restricted in their habitats but occur locally in great abundance. Unlike the pelecypods, however, a greater proportion of the commoner forms are complexly ornamented, and they also possess a much greater potentiality for form variation than do the bivalved mollusks. Consequently, Pennsylvanian gastropods appear to offer a greater opportunity for the discrimination of stratigraphically significant varieties.

The work of J. B. Knight gives some indication of the potentialities of the gastropods, although his studies apparently were not directed toward the recognition of evolutionary series. Some of my own rather casual observations show that such series do exist among the pleurotomarids.

Cephalopods generally are neither abundant enough nor well enough preserved to be good index fossils in the Pennsylvanian. The commoner forms are nautiloids, which largely lack characters that may be used to identify closely discriminated varieties. The ammonoids, which possess much greater possibility in this regard, are mostly rare or very sporadic in their occurrence. In spite of recent studies by F. B. Plummer and Gayle Scott and by A. K. Miller and associates, the stratigraphic ranges of few species seem to be adequately known.

Trilobites are rarely even moderately abundant in the Pennsylvanian, and their remains generally occur in a dismembered condition. Recognized species and varieties also are few. Nevertheless, my studies have proved the existence of evolutionary series in the two principal genera of this age, and trilobites have been successfully used for broader correlations.

CONCLUSIONS

Investigation of the Pennsylvanian faunas of North America has lagged behind studies of the fossils of the other

Paleozoic systems. Because of this and because conspicuous faunal breaks are largely absent in the Pennsylvanian, paleontologic correlations are less precise and less certain than in several older systems. Important studies now in progress will undoubtedly furnish much better evidence for correlation than now exists, but, as long as correlations are made on a percentage basis or by the stratigraphic range-of-species method, only rather general results can be expected, which will mainly confirm rather than refine most of the correlations that are held today. On the other hand, the careful working-out of lines of evolutionary development in a reasonable number of the more common Pennsylvanian fossils is almost certain to furnish the basis for closer and more reliable interregional correlations.

Variation within evolutionary series was fairly slow, however, and there is small likelihood that paleontology can ever compete seriously with correlations based on lithologic peculiarities and the comparison of detailed stratigraphic sequences in individual basins where cyclothemic deposition occurred.

RÔLE OF MICROFOSSILS IN INTERREGIONAL PENNSYLVANIAN CORRELATIONS[1]

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ABSTRACT

In the past the use of microfossils for long-distance correlations in the American Carboniferous has been somewhat restricted, largely because of the few workers interested in this class of fossils. Previous work has centered chiefly on the fusulines and ostracodes, and papers on these fossils have come mostly from the Mid-Continent region, as have also the relatively few papers on other Foraminifera and conodonts. More recent work on microfaunas from formations in the eastern Interior Basin and the West has shown that microfaunas, particularly some Foraminifera and the ostracodes, can be utilized as valuable tools in stratigraphic work.

INTRODUCTION

Three groups of microfossils, namely, the Foraminifera (particularly the fusulines), ostracodes, and conodonts have been employed in Pennsylvanian correlation studies. More than a quarter-century of work, however, has shown that the fusulines are much more useful than any of the others. In the chart and report of the National Research Council on the correlation of Pennsylvanian formations in North America[3] the fusulines are given first rank over other classes of fossils as zone markers in this system. Ostracodes have been little used in this system, largely because of the lack of sufficient work to determine diagnostic species in the several sedimentary basins. However, with the completion of three comprehensive reports—one on the Ardmore Basin, the second on the Kansas-Oklahoma area, and a third on the Illinois Basin—it is now possible to recognize a number of species that are good horizon markers. The work that has been done on Pennsylvanian conodonts has shown that, with few exceptions, the stratigraphic range of these fossils is too long for them to be useful as index fossils. This is also true of the Foraminifera other than fusulines.

Other types of microfossils, such as fragments of holothurians, echinoids, and other fossils, which include dissociated plates and other structural members, have received scant attention. Some work has been done on microcrinoids, but it is insufficient to be of stratigraphic value, except possibly at one or two horizons. Embryonic to youthful forms of the common larger fossils have been given little study or have been neglected entirely. Likewise, the Bryozoa represent, in general, a little-worked group, which probably would be of considerable stratigraphic value if more universally utilized. Some attention has also been given to sponge spicules. Finally, the diagnostic value of such things as scolecodonts, diatoms, radiolaria, molluscan radulae and opercula, and skeletal fragments of arthropods and fish is nil, largely either because of their sparse distribution or because of lack of sufficient data on which to make stratigraphic

[1] Published with permission of the chief, Illinois State Geological Survey.

[2] At the time this paper was prepared, Dr. Cooper was with the Illinois State Geological Survey.

[3] Raymond C. Moore *et al.*, "Correlation of the Pennsylvanian Formations of North America," *Bull. Geol. Soc. Amer.*, Vol. LV (1944), pp. 657–706 and chart.

comparisons. The use of plant material, particularly microspores, will be discussed in another paper.

FORAMINIFERA

The fusulines furnish many of the more useful fossils for the identification of Pennsylvanian formations, and the study of these forms in recent years has become a distinctly specialized branch of paleontology. Because of their structural complexity, relatively large size, and abundance in many marine zones, they make excellent guide fossils. As a result, they have been used extensively in zoning rocks of the Pennsylvanian and Permian systems, not only in North America but in many other parts of the world. A number of important papers have been published on the American Pennsylvanian forms, particularly in the Mid-Continent and contiguous areas. Many long-range correlations have been made with geologic sections outside North America, particularly those in Russia, China, and Japan—in fact, these fossils have been described from all continents.

The stratigraphic range of the Fusulinidae depends upon the classification favored. If those forms which are thought to be ancestral, such as *Endothyra*, etc., are considered true fusulines, the family extends into the Lower Mississippian or possibly even into the Upper Devonian. However, many have considered that the true fusulines are confined to the Pennsylvanian and Permian, and only recently have they been found in the Mississippian (*Millerella* of Illinois, Arkansas, and Wyoming).

The Pennsylvanian fusuline zones are based upon the predominant occurrence of the four genera, *Millerella*, *Fusulinella*, *Fusulina*, and *Triticites*. These zones overlap somewhat but correspond in the main to the groups and series of the system. Many have been further subdivided into subzones, based upon the occurrence of certain species or, in some cases, on certain genera. However, many species are geographically restricted to a single sedimentary province, having reacted to environmental conditions much as other animals do. An excellent discussion of Lower Pennsylvanian fusulines by R. C. Spivey and T. G. Roberts[4] appeared in a recent issue of the *Bulletin of the American Association of Petroleum Geologists*.

Millerella zone.—While this genus ranges from the lowest Pennsylvanian marine formations to as high as the Cisco of Texas, the *Millerella* zone has been designated by M. L. Thompson[5] as embracing those beds in which this genus occurs to the exclusion of other fusulines of the normal long-axis type. The zone is known to be well marked in Kansas, Arkansas, Oklahoma, Texas, New Mexico,[6] Utah, and Colorado;[7] and, no doubt, further work will show it to be present in other areas where marine rocks of Springer (Morrow) age are developed.

Fusulinella zone.—This genus characterizes the section above the Springer, containing beds of Atoka and lowest Des Moines age. It is present in the Pottsville (Mercer) limestones of Ohio, the Upper Marble Falls of Texas, the Tradewater

[4] "Lower Pennsylvanian Terminology in Central Texas," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XXX (1946), pp. 181–86.

[5] "Pennsylvanian Morrowan Rock and Fusulinids of Kansas," *Kan. Geol. Surv. Bull. 52*, Part VII (1944), pp. 409–31, Pls. 1, 2, Figs. 1, 2.

[6] M. L. Thompson, "Pennsylvanian System in New Mexico," *N.M. School of Mines Bull. 17* (1942).

[7] M. L. Thompson, "Pennsylvanian Rocks and Fusulinids of East Utah and Northwest Colorado Correlated with Kansas Section," *Kan. Geol. Surv. Bull. 60*, Part II (1945), pp. 17–84, Pls. 1–6, Figs. 1–11.

(Seville) of Illinois, and the lowermost Des Moines of Iowa. A lower subzone, characterized by *Profusulinella* (or *Fusiella* of Lee), and an upper zone, the *Fusulinella iowensis* subzone, are present. Only recently the suggestion has been made that a group of closely related species of *Fusulinella*,[8] which are characteristic of pre–Des Moines beds, be considered the "zone of *Fusulinella*," even though younger species are associated with *Wedekindellina* in higher zones.[9] Other genera commonly found associated with *Fusulinella* are *Millerella*, *Pseudostaffella*, and *Eoschubertella*.

Fusulina zone.—This zone, immediately above the *Fusulinella* zone, delineates all the Des Moines not included in the lower zone. Its upper boundary, the Des Moines–Missouri line, is the position of the most significant faunal break in the Pennsylvanian, not only for the fusulines but for many other classes of fossils as well. Although *Fusulina* is found with *Fusulinella* in the lowest Des Moines, it does not occur above the upper boundary of this series. Subzones are marked by *Wedekindellina* in the lower part and *Fusulina girtyi* and *F. exima* in the upper Des Moines. These zones are well marked in the Mid-Continent, Rocky Mountain, and central Interior basins. Only the upper subzone is known in Ohio, *Wedekindellina* apparently being absent there.[10]

Triticites zone.—Just as *Fusulina* is confined to beds below the Des Moines–Missouri boundary, *Triticites* is not found in the formations below the Missouri series of the Upper Pennsylvanian. However, it continues upward into the Lower Permian (Wolfcamp series). The zone is widespread in North America, and three zones, characterized by *T. irregularis*, *T. cullomensis*, and *T. ventricosus*, are recognized. In the lower zone *T. irregularis* is similar to *T. ohioensis* and appears to occur at about the horizon of the Kansas City group. Other genera, such as *Millerella*, *Schubertella*, *Waeringella*, and *Dunbarinella* make up only minor parts of the fusuline faunas of the Upper Pennsylvanian.[11] An excellent bibliography and systematic review of genera was published by Dunbar and Henbest.[12]

The smaller Foraminifera have not furnished horizon markers in the same proportion as have the fusulines. This is partly due to the lack of extensive work on these forms but it is, in the main, due to the long range of most of the known genera and species, again reflecting the reappearance of like forms with the repetition of similar environmental conditions. Mrs. Plummer[13] has found, after her extensive work on the Texas section, that "most of the species have such long vertical ranges that they are of little value in this depositional province. Faunal assemblages apparently follow facies, and, with no major breaks in the sequence, faunas in general are repeated with repetition of facies in the column." In writing about the Marble Falls Foraminifera, she says:

> In this thick and almost consistently calcareous Marble Falls formation, a thin incidental noncalcareous and sandy layer is characterized wholly by the same forms that are so highly characteristic of superjacent noncalcareous and silty to sandy formations, and it emphasizes forcibly the need to study faunal groups in close association with obvious field

[8] Thompson, p. 29 of ftn. 6.

[9] Spivey and Roberts, pp. 182–85 of ftn. 4.

[10] Carl O. Dunbar and Lloyd G. Henbest, "Pennsylvanian Fusulinidae of Illinois," *Ill. Geol. Surv. Bull. 67* (1942), p. 30.

[11] Thompson, p. 42 of ftn. 7.

[12] Ftn. 10.

[13] Personal communication.

TABLE 1*

SUMMARY FAUNAL CHART SHOWING FORAMINIFERA FROM LOWER PENNSYLVANIAN OF TEXAS

SPECIES	Marble Falls	Smithwick	Strawn	Unrestricted
Proteonina cervicifera Cushman and Waters				x
Thurammina texana Cushman and Waters				x
Thuramminoides sphaeroidalis, n. sp.				x
Hyperammina bulbosa Cushman and Waters				x
Hyperammina clavacoidea, n. sp.				x
Hyperammina elegantissima, n. sp.				x
Hyperammina spinescens Cushman and Waters				x
Hyperamminoides expansus, n. sp.				x
Earlandia minuta (Cushman and Waters)				x
Reophax arenatus (Cushman and Waters)			x	
Reophax asper Cushman and Waters				x
Reophax bendensis, n. sp.	x			
Reophax emaciatus, n. sp.				x
Reophax expatiatus, n. sp.			x	
Reophax fittsi (Warthin)				x
Reophax glennensis (Harlton)				x
Reophax minutissimus, n. sp.				x
Reophax tumidulus, n. sp.			x	
Ammodiscus semiconstrictus Waters				x
Glomospira compressa Waters		x		
Glomospira articulosa, n. sp.				x
Glomospirella umbilicata (Cushman and Waters)				x
Haplophragmoides confragosus, n. sp.				x
Ammobaculites minutus Waters				x
Ammobaculites stenomecus Cushman and Waters				x
Endothyra distensa, n. sp.	x			
Endothyra rotaliformis Warthin	x			
Endothyra whitesidei Galloway and Ryniker				x
Endothyranella armstrongi sobrina, n. subsp.	x			
Bradyina holdenvillensis Harlton	x			
Bradyina sp.	x			
Bigenerina perexigua, n. sp	x			
Cribrostomum marblense, n. sp.	x			
Cornuspira sp.	x			
Calcitornella sp.	x			
Trepeilopsis sp.	x			
Globivalvulina sp.	x			
Mooreinella biserialis Cushman and Waters				x
Polytaxis scutella (Cushman and Waters)	x			

*After Helen J. Plummer (see ftn. 14).

relationships and with lithology of the enclosing strata, in order not to be misled by species.[14]

The available literature is very limited in scope and is published for the most part under the names of a small number of workers—namely, J. A. Cushman and J. A. Waters,[15] J. J. Galloway and B. H. Harlton,[16] J. J. Galloway and Charles Rynicker,[17] B. H. Harlton,[18] Helen J. Plummer,[19] Robert Roth and John Skinner,[20] A. S. Warthin,[21] and J. A. Waters.[22] Therefore, it is not possible, at the present state of our knowledge, to make comparisons between the smaller foraminiferal faunas of the various sedimentary basins of North America. My own information on the Illinois Pennsylvanian does not permit conclusions that are in any way different from those reached by Mrs. Plummer from her study of the Texas section.

Of the previously described species shown in Table 1, five have been found restricted to a single formation, three to the Marble Falls, and one each to the Smithwick and Strawn. In her analysis of the Texas faunas, Mrs. Plummer reports the new species *Hyperammina elegantissima*, *Reophax emaciatus*, *R. expatiatus*, and *R. tumidulus* as good markers for the Strawn, and *R. bendensis* for the Marble Falls. No restricted species were listed from the Smithwick, all species ranging upward into the Strawn.

OSTRACODES

There is a considerable volume of literature on the Pennsylvanian ostracodes of North America, consisting mostly of short papers on faunas that are very limited stratigraphically. In order to obtain a comprehensive stratigraphic summary it is necessary to examine this extensive literature describing many hundreds of species from scores of localities and formations, principally from the Mid-Continent and adjacent areas. The most important papers are by H. H. Bradfield,[23]

[14] Helen Jeanne Plummer, "Smaller Foraminifera in the Marble Falls, Smithwick, and Lower Strawn Strata around the Llano Uplift in Texas," *Tex. Univ. Pub. 4401* (1945), p. 215.

[15] "Pennsylvanian Foraminifera from Michigan," *Contr. Cushman Lab. Foram. Res.*, Vol. III (1927), pp. 107–10, Pl. 22; "Arenaceous Paleozoic Foraminifera from Texas," *ibid.*, pp. 146–53, Pls. 26, 27; "The Development of *Climacammina* and Its Allies in the Pennsylvanian of Texas," *Jour. Paleon.*, Vol. II (1928), pp. 119–30, Pls. 17–20; "Some Foraminifera from the Pennsylvanian and Permian of Texas," *Contr. Cushman Lab. Foram. Res.*, Vol. IV (1928), pp. 31–35, Pls. 4–7; "Additional Cisco Foraminifera from Texas," *ibid.*, pp. 62–67, Pl. 8; "Upper Paleozoic Foraminifera from Texas," *Jour. Paleon.*, Vol. II (1928), pp. 358–71, Pls. 47–49; "Foraminifera from the Cisco Group of Texas," *Tex. Univ. Pub. 3019* (1930), pp. 22–81, Pls. 2–12.

[16] "Some Pennsylvanian Foraminifera of Oklahoma, with Special Reference to the Genus *Orobias*," *Jour. Paleon.*, Vol. II (1928), pp. 338–57, Pls. 45, 46; "*Endothyranella*, a Genus of Carboniferous Foraminifera," *ibid.*, Vol. IV (1930), pp. 24–28.

[17] "Foraminifera from the Atoka Formation of Oklahoma," *Okla. Geol. Surv. Circ. 21* (1930).

[18] "Some Pennsylvanian Foraminifera of the Glenn Formation of Southern Oklahoma," *Jour. Paleon.*, Vol. I (1927), pp. 15–27, Pls. 1–5; "Pennsylvanian Foraminifera of Oklahoma and Texas," *ibid.*, pp. 305–10, Pls. 52, 53; "Micropaleontology of the Pennsylvanian Johns Valley Shale of the Ouachita Mountains, Oklahoma, and Its Relationship to the Mississippian Caney Shale," *ibid.*, Vol. VII (1933), pp. 3–29, Pls. 1–7.

[19] "Calcareous Foraminifera in the Brownwood Shale near Bridgeport, Texas," *Tex. Univ. Pub. 3019* (1930), pp. 1–21, Pl. 1.

[20] "The Fauna of the McCoy Formation, Pennsylvanian, of Colorado," *Jour. Paleon.*, Vol. IV (1930), pp. 332–52, Pls. 28–31.

[21] "Micropaleontology of the Wetumka, Wewoka, and Holdenville Formations," *Okla. Geol. Surv. Bull. 53* (1930).

[22] "A Group of Foraminifera from the Dornick Hills Formation of the Ardmore Basin (Oklahoma)," *Jour. Paleon.*, Vol. I (1927), pp. 129–33, Pl. 22; "A Group of Foraminifera from the Canyon Division of the Pennsylvanian Formations in Texas," *ibid.*, Vol. I (1928), pp. 271–75, Pl. 42.

[23] "Pennsylvanian Ostracoda of the Ardmore Basin, Oklahoma," *Bull. Amer. Paleon.*, Vol. XXII, No. 73 (1935).

Coryell *et al.*,[24] B. H. Harlton,[25] W. R. Johnson,[26] and Betty Kellett.[27] Of these, two stand out because of their comprehensive study of ostracodes from a large stratigraphic section. Bradfield's report on the ostracodes of the Ardmore Basin, Oklahoma, describes the fauna from all the Pennsylvanian overlying the Springer formation. This section embraces some 3,000 feet or more of beds in the Dornick Hills, Deese, and Hoxbar groups, which range in age from Morrow to the top of the Missouri series. Kellett's reports involve the Pennsylvanian and Permian section of Oklahoma and Kansas and give the stratigraphic distribution of species in formations ranging from the base of the Missouri to the top of the Big Blue (Permian) series. The two reports overlap, in that each describes ostracodes from the Missouri series.

An analysis of the range of Paleozoic ostracode genera made a few years ago[28] showed that over half the Mississippian genera did not continue into the Pennsylvanian. In this category might be mentioned a few characteristic genera which are present in the Chester, such as *Bairdiolites*, *Carboprimitia*, *Chesterella*, *Denisonella*, *Graphiodactylis*, and *Perprimitia*. A like portion of the Pennsylvanian genera originate in the Mississippian or earlier, and, of these, about two-thirds continue into the Permian.

The results of my own studies on the Pennsylvanian of Illinois[29] give the ostracode faunas from zones ranging in age from Morrow to the top of the Virgil, representing the entire Mid-Continent section, except for some horizons present west of the Mississippi that are missing or unfossiliferous in Illinois. An analysis of all the available data shows that, when the total number of species is considered, comparatively few species are good horizon markers.

[24] H. N. Coryell, "Some New Pennsylvanian Ostracoda," *Jour. Paleon.*, Vol. II (1928), pp. 87–94, Pl. 11, pp. 377–79, Pl. 51; H. N. Coryell and G. A. Osorio, "Pennsylvanian Ostracoda, and Ostracode Fauna from the Nowata Shale," *Amer. Mid. Nat.*, Vol. XIII (1932), pp. 25–40, Pl. 5; H. N. Coryell and Gladys D. Billings, "Pennsylvanian Ostracoda of the Wayland Shale of Texas," *Amer. Mid. Nat.*, Vol. XIII (1932), pp. 170–89; H. N. Coryell and C. H. Sample, "Pennsylvanian Ostracoda: A Study of the Ostracode Fauna of the East Mountain Shale, Mineral Wells Formation, Texas," *Amer. Mid. Nat.*, Vol. XIII (1932), pp. 245–81, Pls. 24–26; H. N. Coryell and R. T. Booth, "Pennsylvanian Ostracoda: A Continuation of the Ostracoda Fauna from the Wayland Shale, Graham, Texas," *Amer. Mid. Nat.*, Vol. XIV (1933), pp. 258–78, Pls. 3–5.

[25] "Some Pennsylvanian Ostracoda of the Glenn and Hoxbar Formations of Southern Oklahoma and the Upper Part of the Cisco Formation of Northern Texas," *Jour. Paleon.*, Vol. I (1927), pp. 203–12, Pls. 32, 33; "Pennsylvanian Ostracodes from Oklahoma and Texas," *ibid.*, Vol. II (1928), pp. 132–41, Pl. 21; "Pennsylvanian Ostracoda from Menard County, Texas," *Tex. Univ. Pub. 2901* (1929), pp. 139–61, Pls. 1–4; "Some Upper Mississippian (Fayetteville) and Lower Pennsylvanian (Wapanucka-Morrow) Ostracoda of Oklahoma and Arkansas," *Amer. Jour. Sci.*, Vol. XVIII, No. 105 (5th ser., 1929), pp. 254–70, Pls. 1, 2; see also ftn. 18 (1933).

[26] "The Ostracoda of the Missouri Series in Nebraska," *Neb. Geol. Surv. Paper 11* (1936), pp. 1–52, Pls. 1–5.

[27] "The Ostracode Genus *Hollinella:* Expansion of the Genus and Description of Some Carboniferous Species," *Jour. Paleon.*, Vol. III (1929), pp. 196–217, Pls. 25–26; "Ostracodes of the Upper Pennsylvanian and Lower Permian Strata of Kansas. I. The Aparchitidae, Beyrichiidae, Glyptopleuridae, Kloedenellidae, Kirkbyidae, and Youngiellidae," *ibid.*, Vol. VII (1933), pp. 59–108, Pls. 13–16; "Ostrocodes from the Upper Pennsylvanian and Lower Permian Strata of Kansas. II. The Genus *Bairdia*," *ibid.*, Vol. VIII (1934), pp. 108–38, Pls. 14–19; "Ostracodes of the Upper Pennsylvanian and Lower Permian Strata of Kansas. III. Bairdiidae (*concl.*), Cytherellidae, Cyprinidae, Entomochonchidae, Cytheridae, and Cypridae," *ibid.*, Vol. IX (1935), pp. 132–66, Pls. 16–18; "Carboniferous Ostracodes," *ibid.*, Vol. X (1936), pp. 769–84.

[28] Chalmer L. Cooper, "Occurrence and Stratigraphic Distribution of Paleozoic Ostracodes," *Jour. Paleon.*, Vol. XVI, No. 6 (1942), pp. 764–76.

[29] "Pennsylvanian Ostracodes of Illinois," *Ill. Geol. Surv. Bull. 70* (1946).

CHART SHOWING PENNSYLVANIAN FAUNAL ZONES

AGE	APPALACHIANS		MIDCONTINENT AND CENTRAL INTERIOR BASIN								
	SER.	FLORAL ZONES	SER.	GROUPS	BRACHIOPODS	AMMONITES	CRINOIDS	FUSULINES ZONE	FUSULINES SUBZONE	OSTRACODES	CONODONTS
PERMIAN	DUNKARD	Callipteris	WOLFCAMP	ADMIRE		Perrinites		Schwagerina	S. emaciata S. huecoensis	Kellettina robusta	
PENNSYLVANIAN	MONONGAHELA	Odontopteris	VIRGIL	WABAUNSEE	Lissochonetes,	Uddenites	Delocrinus vulgatus	Triticites	T. ventricosus	Bairdia concava	Streptognathodus elongatus
		Danaeites		SHAWNEE	Chonetes transversalis				T cullomensis	Hollinella moorei	Prioniodus? conflexus
		Lescuropteris		DOUGLAS							
	CONEMAUGH	Pecopteris	MISSOURI	PEDEE LANSING	Chonetina	Eothallaoceras	Apographiocrinus typicalis			Hollinella radlerae	Gondolella spp
				KANSAS CITY		Prouddenites			T. irregularis	Bairdiacypris deloi	
				BRONSON						Bairdia spinosa	
				BOURBON							
	ALLEGHENY	Neuropteris flexuosa	DES MOINES	MARMATON	Mesolobus mesolobus	Owenoceras	Delocrinus granulosus	Fusulina	F. exima F. girtyi	Silenites lenticularis	Gondolella magna G. cruvata
		Neuropteris rarinervis		CHEROKEE	Marginifera muricatina	Wellerites	Plaxocrinus mooresi		Wedekindellina	Cavellina equalis	Idiognathus acutus
	KANAWHA	Neuropteris tenuifolia	TI VALLEY	ATOKA	Marginifera nana M. haydenensis Mesolobus striatus	Paralegoceras		Fusulinella	F. iowensis	Bairdia ardmorensis	Streptognathodus n. sp.
		Cannophyllites									
	LEE	Mariopteris pygmaea							Profusulinella		
		M. pottsvillea Aneimites		MORROW (Springer)	Hustedia miseri	Gastrioceras	Stereobranchi-crinus	Millerella	M. marblensis	Polytylites wapanuckensis	Polygnathus wapanuckensis
		N. pocahontas M. eremopteroides				Branneroceras			M. inflecta		Streptognathodus ouachitensis
S.	CH NK									Bairdiolites	Cavusgnathus

Ti Valley series.—This name is suggested in the Illinois report as a solution for the Morrow-Atoka-Lampasas–Marble Falls nomenclatural problem and includes the Morrow and Atoka groups, the latter being used with the stratigraphic connotation of Spivey and Roberts.[30] The series is characterized by *Ectodemites harltoni*, from the Johns Valley shale and the Ferdinand of southern Indiana, and *Kirkbya bendensis*, from the same zones. The lower part of the series, the Morrow, contains the distinctive *Polytylites wapanuckensis* as well as *Monoceratina ardmorensis*. In the Atoka group the lower Ferdinand formation contains *Bairdia ardmorensis* and *Ectodemites plummeri* from the Dornick Hills and Marble Falls, respectively. The next higher formation, the Seville, does not have the restricted species common to zones in the Mid-Continent, but several new species appear to be restricted in the Illinois Basin.

Des Moines series.—In the Mid-Continent and Illinois Basin areas this series provides a number of species common to both groups and formations. *Cavellina equalis* is found in the lowest group, the Cherokee, and *Coryellites cooki* is characteristic of the Liverpool and upper Deese.

The Marmaton group contains *Geisina gregaria*, *Microparaparchites brazoensis*, and *Silenites lenticularis*. The Summum contains *Healdiacypris perplexa* from the upper Deese and the Brereton, while *Geisina gallowayi*, *Jonesina dubia*, and *Bairdia citriformis* are from comparable horizons in the west. The position of the Lonsdale near the top of the Des Moines is indicated by such species as *Cavellina cavellinoides*, *C. lintris*, *Hollinella nowataensis*, *Jonesina deesensis*, and *Microparaparchites cuneatus*. The next higher formation, the Exline, contains *Cavellina jejuna*, *Healdia formosa*, and *Jonesina trisulcata*.

Missouri series.—*Kirkbya arcuata* marks this series, but no species are known to be restricted to the oldest group, the Bronson. The Kansas City is indicated by *Bairdiacypris deloi* and *Jonesina infrequens*. In the Lansing group the La Salle–Livingston-Millersville zone is characterized by *Hollinella radlerae*, *Bairdia hooverae*, *B. lunata*, *Ectodemites sullivanensis*, and *Bairdiacypris haydenbranchensis*, the first three from the upper Missouri of the Mid-Continent and the other two from Indiana.

Virgil series.—The Virgil contains *Bairdia scholli* and *Cavellina laevis*, which also are found in the upper formations of the Illinois section. The Shawnee Little Vermilion formation contains *Healdia masoni* and *Knightina ampla*, while the Newton-Shumway zone has *Bairdia pinnula*, *Cavellina nebrascensis*, *Healdia coryelli*, *Hollinella moorei*, and *Silenites silenus*. The youngest Illinois formation, the Woodbury, is marked by *Ectodemites edsonae*, present in the Hoxbar of Oklahoma, somewhat lower in the section. Kellett's work has shown that "almost all of the Permian ostracode genera are found also in late Pennsylvanian strata, and with few exceptions, the same genera predominate. Many of the common species in earliest Permian faunas are Pennsylvanian survivals, but in addition a dozen new species occur. By mid-Permian (Leonard) time few Pennsylvanian species survived, although a number of close relatives may be recognized. In general mid-Permian ostracodes are more highly ornamented than Pennsylvanian species, especially in the families Kirkbyidae and Kloedenellidae."[31]

[30] P. 185 of ftn. 4; see also Cooper, ftn. 30.

[31] Personal communication.

CONODONTS

Conodonts, so numerous and varied in the Upper Devonian and Lower Mississippian, while still rather plentiful in the Upper Mississippian and Pennsylvanian, are represented by few genera and species in the younger system. They are quite numerous in the black shales, which have furnished the many assemblages that have been described. The literature contains comparatively few papers on Pennsylvanian conodonts, the principal workers being Samuel Ellison,[32] F. H. Gunnell,[33] and C. R. Stauffer and Helen J. Plummer.[34] Like many other classes of fossils, conodonts are creatures of environment and so, like many other Pennsylvanian forms, are largely of long stratigraphic range. The most important stratigraphic and taxonomic study[35] shows that nearly all species lived throughout Pennsylvanian times. One species, *Cavusgnathus giganta*, apparently started in the Chester and lived on into the Permian. Very little is known about conodonts from the Upper Mississippian or from those formations classed as "Mississippian and/or Pennsylvanian." As in other transition zones, genera from both systems are found, such as *Cavusgnathus*, *Gnathodus*, and *Metalonchodina*. The following details of the fauna and the chart are summarized from Ellison's report:[36] Species of *Prioniodus*, *Ligonodina*, *Lonchodina*, *Hibbardella*, *Prioniodina*, *Synprioniodina*, *Ozarkodina*, *Spathodus*, and *Hindeodella* are present throughout the system, some being restricted, many extending into the Permian.

Ti Valley series.—The only described fauna from this series is that of Harlton,[37] and this contains only three species. Two features distinguish this series, namely, *Polygnathodella* is known only from the Morrow, and *Gondodella* is not known to range below the base of the Des Moines.

Des Moines series.—The most important feature of this series is the presence of large numbers of idiognathodids, but the only restricted species is *Idiognathodus acutus*. *Gnathodus* and *Metalonchodina* are also rather abundant and are not known in younger formations. While *Gondodella* is not common, *G. curvata* and *G. magna* are characteristic of several zones within the Marmaton.

Missouri series.—Most striking is the abundance of idiognathodids, streptognathodids, and gondodellids: *I. lobatus*, *Streptognathodus sulcatus*, *S. concentricus*, and *S. oppletus* are confined to the series. There is a noticeable disappearance of many idiognathodids near the top of the Missouri. *Gondodella dubia* is confined to the Hushpuckney member of the Swope formation, and *G. symmetrica* to the Stark shale member of the Dennis, both in the Bourbon group. Two species of *Cavusgnathus*—*C. lauta* and *C. flexa*—begin in the Missouri but carry on into the Permian.

Virgil series. — Streptognathodids, mostly *S. elegantulus*, make up over three-fourths of the Virgil fauna. Here *S. elegantulus* and *Prioniodus conflexus* appear for the first time and continue

[32] "Revision of Pennsylvanian Conodonts," *Jour. Paleon.*, Vol. XV (1941), pp. 107–43, Pls. 20–23; "Conodonts as Paleozoic Guide Fossils," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XXX, No. 1 (1946), 93–110.

[33] "Conodonts from the Fort Scott Limestone of Missouri," *Jour. Paleon.*, Vol. V (1931), pp. 244–52, Pl. 29.

[34] "Texas Pennsylvanian Conodonts and Their Stratigraphic Relations," *Tex. Univ. Bull. 3201* (1932), pp. 13–50.

[35] Ellison, ftn. 33 (1941).

[36] *Ibid.*, pp. 108–11, Fig. 4.

[37] See ftn. 18 (1933).

into the Big Blue series. All gondodellids except *Gondodella denuda* and *G. elegantula* disappear in zones above the lower Virgil Lecompton formation.

This analysis of conodont faunas pertains to the Pennsylvanian of Missouri and Kansas. There are no comparable faunal data covering other sedimentary basins. Outside of a few papers on Oklahoma formations by Harlton[38] and by R. W. Harris and R. V. Hollingsworth,[39] the only other comprehensive Pennsylvanian conodont paper is by Stauffer and Plummer[40] on the Texas Pennsylvanian. My own observation is that there is little variation in conodont faunas throughout the whole system or from one sedimentary basin to another. They seem to follow ecological conditions closely, and faunas are found to be repeated numerous times under the multiple repetition of similar beds during this system. The few possibilities for guide fossils are the platform types, which have furnished the small number of restricted forms that are now known. Even these may possibly be shown to have considerable range by future work. More reports like that by Ellison are needed for other areas.

[38] *Ibid.*

[39] "New Pennsylvanian Conodonts from Oklahoma," *Amer. Jour. Sci.*, Vol. XXV (5th ser., 1933), pp. 193–204, Pl. 1.

[40] See ftn. 34.

PENNSYLVANIAN FLORAL ZONES AND FLORAL PROVINCES[1]

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ABSTRACT

Paleontologic correlation must be guided by principles, some of which are stated here, derived from a study of living organisms and their distribution. In application of these principles, nine floral zones in the Arcto-Carboniferous province of Pennsylvanian floras are considered. The floras included are all lowland, warm-climate, rain forest assemblages. In the Rocky Mountain area certain floral types appear that are thought to represent upland floras. In the Antarcto-Carboniferous province the Pennsylvanian floras differ greatly in the main and are believed to represent coal-climate, rain forest assemblages. A few representatives of these assemblages are considered, and a few Arcto-Carboniferous elements present are regarded as immigrants during an epoch of milder climate.

INTRODUCTION

The following statement, written in connection with a discussion of Devonian faunas over forty years ago, requires only slight modification to make it applicable to land floras. It contains, in fact, many of the fundamental ideas of sound stratigraphic paleontology. The import of these remarks has frequently been lost sight of but is of great consequence.

> Students of geographical distribution have shown that in distant parts of the same ocean the species are widely divergent, as much difference existing between the marine faunas of the southern and northern temperate zones as between the faunas of the successive formations of a continuous geological section. It is evident from this observation that discussions of the time relations of fossils must treat not only of the genetic affinity of the forms making up a fauna, but of the geographic distribution and of the geological range of the species concerned.[2]

As a basis for the ensuing presentation of paleontologic data, it is pertinent to introduce several of the principles of the science that are of value in the study of Paleozoic floras. Their statement may recall them to more active use, with beneficial results, and it is with this purpose that they are set down.

First is the biological principle that forms the fundamental basis for paleontologic work: The facts and accepted hypotheses of modern biology, although not always applicable, are the bases for studies of fossil organisms. The features of modern classifications that are proved to be of importance must be the elements of any classification of fossils. It is generally recognized that the paleontologist, owing to imperfect preservation of the remains with which he works, must accept, as his primary bases for determining identity and relationship, superficial characteristics that are deemed of secondary importance in the classification of most living organisms. This fact he should fully admit, and he should recognize that, in consequence, superficial homeomorphy may be wrongly interpreted and that many patterns of "evolutionary trends" may be more apparent than real. In short, the paleontologist must be fully aware, and be prepared to admit, that interpretations of generic and even specific relationships of fossil organisms are based on opinion rather

[1] Published by permission of the director, U.S. Geological Survey.

[2] H. S. Williams, "The Correlation of Geological Faunas," *U.S. Geol. Surv. Bull. 210* (1903), p. 15.

than on the biologic details that permit the student of living organisms to determine affinity.

Several ecological and geographical principles are of considerable importance in stratigraphic work. They are, briefly, as follows:

1. *The fundamental unit in ecologic and geographic studies is the biota or life-association.*—The biota is also fundamental in geologic work. Since the paleontologist rarely has recourse to the complete association, he should accept the largest and most representative sample available. Should he choose to make a special study of only a part of the sample, such as a family or a genus, he should admit this limitation by qualification of his results.

2. *Organisms are rarely cosmopolitan.*—There are very few known forms so adaptable that they have a general or universal distribution on continental masses.

3. *Associations living during a limited period of time in the same general region are much more alike in their constitution and the proportionate abundance of dominant species than are those of widely separated regions.*—In this principle we find a justification for the method of correlation in restricted areas through abundance of certain species in a stratal zone and an explanation of the common failure of such a zone to maintain its identity over wide areas. We likewise find an explanation of variation in fossil content in widely separated areas of strata that we believe to be contemporaneous.

4. *The association of elements of one flora with those of another suggests an intermediate environment if, on other bases, the floras are known to represent different habitats.*—If they are considered to represent similar or identical habitats, such floras are indicative of different stratigraphic positions, or they may be from some intermediate region that marks the general position of a path of intercommunication or migration.

5. *Variations in content of associations may be due to geographic distribution or geological range.*

6. *Widespread distribution of organisms or associations is to be expected in a negligible period of time only when there is adequate evidence that the areas occupied are equally and broadly continuous. If the areas occupied are recognized as discontinuous, it is then to be expected that the geologic time involved will be variable but not always negligible.*

7. *Identical genera or species may frequently be traced through several life-associations that form a geographic or a geologic succession or both. Accepting correlation by the use of associations, the oldest occurrence of a species or genus suggests the approximate locus of its origin.*

8. *Of paramount importance in geographic and geologic work is the recognition of the facts that* (a) *practically simultaneous changes in forms often occur over areas of great geographic spread and* (b) *more protracted changes in forms occur in a small area during appreciable intervals of geologic time.*—These changes may develop uniformly, or they may take place in surges. Stratigraphic variation in species, genera, or biotas may be explained by migration, changes, or hiatuses. A hiatus can be recognized, on paleontologic grounds, with certainty and properly evaluated only when the suspected local absence of a biota can be proved through its presence elsewhere in the region.

Further, I wish to present three opinions concerning associations of Pennsylvanian floras:

1. Floras of the Pennsylvanian coal swamps of the Northern Hemisphere

have many morphological characteristics that suggest that they persisted under conditions no cooler than those of present-day, warm-temperate rain forests.

2. Floras of the Pennsylvanian equivalents in the Southern Hemisphere originated under conditions no warmer than, and possibly similar to, the cool-temperate rain forests or provinces of the present day.

3. Upland floras are believed to have existed during much, if not all, of the period of existence of land plants. Their remains are not likely to be preserved in the fossil record because of distance of habitats from basins of deposition. They may be expected in the sedimentary record if local conditions were such that basins occurred adjacent to uplands.

FLORAS OF PENNSYLVANIAN AGE

Two major floral provinces have been recognized as existing during the Pennsylvanian epoch: (1) the northern or Arcto-Carboniferous province of warm-climate, rain forest coal floras and (2) the southern or Antarcto-Carboniferous province of cool-climate, rain forest floras.

ARCTO-CARBONIFEROUS FLORAS

STANDARD SEQUENCE OF PENNSYLVANIAN FLORAS

The study of Pennsylvanian fossil plants has been largely a study of the "Coal Measures" floras, groups of plants that occur in strata that are largely continental in origin and that are believed to have accumulated in extensive flood plains and in deltas under mild and humid climatic conditions. Nine successive floral zones are recognized in the eastern and Mid-Continent regions of North America. These zones provide a practical basis for general correlation of the containing rocks, "general correlation" not implying the precise establishment of boundaries of the floral zones. Obviously, in thick sequences of clastic strata the possibility of preservation everywhere of characteristic floras at the precise position of their extinction is unlikely. The stratigrapher can, therefore, only "rough in" the stratigraphic units through strict paleontologic work.

Secondary floral characteristics, such as relative abundance of certain species, may be further used for detailed local correlations, but only rarely are they valuable over large areas. Preferred for refinements of correlation within floral zones when possible are field mapping and direct tracing.

The nine floral zones have been named from characteristic genera or species. Such naming of zones does not mean, however, that knowledge of the index or marker fossils alone is sufficient for recognition. The names given are those of certain more common forms; but definite zone identification requires a study of the entire flora when possible. The nine zones and their gross distribution are given in Table 1. The following paragraphs discuss these zones in greater detail.

Zone 1.—Zone of *Neuropteris pocahontas* and *Mariopteris eremopteroides*. This zone is recognized as characteristic of the lower Lykins coals of the Pottsville formation in the Anthracite fields of Pennsylvania and of the Pocahontas formation in West Virginia. It is likewise identifiable in the southern Appalachians in areas of expansion of the Pottsville formation. Throughout much of the area the limits of the zone have not been clearly established. Correlative units probably exist in many parts of the United States but are rarely of facies suitable for preservation of fossil plants.

Zone 2.—Zone of *Mariopteris pottsvillea* and of common occurrence of *Aneimites* spp. Floras belonging to this

zone occur in Lykins coal No. 4, in the lower part of the New River formation of West Virginia, and at traceable positions farther south. Such floras also occur locally at the base of the Pennsylvanian series in the Mid-Continent region. Thus the Wayside member of the Caseyville formation, southern Illinois, carries this association.

TABLE 1

Zone	Name	Appalachian Region	Mid-Continent Region
9.....	*Danaeites*	Upper part of Monongahela formation	In Mid-Continent region Zones 8 and 9 are not separable and could together be designated the zone of *Odontopteris* spp.
8.....	*Lescuropteris*	Lower part of Monongahela formation and upper part of Conemaugh formation	
7.....	*Neuropteris flexuosa*, and *Pecopteris* spp.	Upper part of Allegheny formation	Upper part of Des Moines group
6.....	*N. rarinervis*	Lower part of Allegheny formation	Lower part of Des Moines group
5.....	*N. tenuifolia*	Major portion of Kanawha formation	Major portion of Lampasas group
4.....	*Cannophyllites*	Base of Kanawha formation	Base of Lampasas group
3.....	*Mariopteris pygmaea*, *Neuropteris tennesseeana*, *Ovopteris communis*, *Alloiopteris inaequilateralis*, and *Alethopteris decurrens*	Coals 2 and 3 of the type Pottsville formation, upper part of New River formation, upper part of Lee formation	Morrow formation
2.....	*Mariopteris pottsvillea* and *Aneimites* spp.	Lykins coal No. 4 of Pottsville formation, lower part of New River formation	Base of Pennsylvanian series in Mid-Continent region
1.....	*Neuropteris pocahontas* and *Mariopteris eremopteroides*	Lower Lykins coals of Pottsville formation and Pocahontas formation	

Zone 3.—Zone of *Neuropteris tennesseeana*, *Ovopteris communis*, *Alloiopteris inaequilateralis*, *Alethopteris decurrens*, and *Mariopteris pygmaea*. As indicated by its contained floras, this zone is widespread in the Appalachian region and in the Mid-Continent area. It occurs in the roofs of coal Nos. 2 and 3 of the type Pottsville, is characteristic of the upper part of the New River formation in West Virginia, and is found in the upper part of the Lee formation farther south. It is likewise known from the Morrow formation and its equivalents in the Mid-Continent region.

The following list of selected plants has been made to illustrate a characteristic flora of this zone. Drurey shale (Battery Rock member), NW. $\frac{1}{4}$ sec. 32, T. 10 S., R. 1 W., Carbondale Quadrangle, Illinois: *Pecopteris serrulata*, *Alethopteris lonchitica*, *A. decurrens*, *A. owenii*, *A. owenii* var. *grandifolia*, *A. owenii* var. *helenae*, *A.* sp., *Neuropteris tennesseeana*, *Neuropteris* spp., *Linopteris* sp., *Mariopteris pygmaea*, *M. speciosa*, *Mariopteris* spp., *Diplothmema cheathami*, *Diplothmema* sp., *Eremopteris inaequilateralis*, *Cardiocarpon* spp., *Cordaites principalis*,

Sphenophyllum sp. cf. *S. longifolium*, *S. cuneifolium*, *Lepidophyllum campbelleanum*, *Sigillariostrobus* spp.

Zone 4.—Zone of common occurrence of *Cannophyllites*. This zone may be recognized at or near the base of the Kanawha formation and its equivalents in the Appalachian region and at the base of the Lampasas group in the mid-continent region. It is frequently difficult to identify because of similarities of many of its elements to those of adjacent strata.

The following list has been compiled from several collections in western Illinois and illustrates the general characteristics of floras assigned to this zone. Tarter member, Tradewater formation in (1) SW. ¼ SW. ¼ sec. 27, T. 11 S., R. 4 E., Marian Quadrangle, Illinois; (2) NE. ¼ SW. ¼ SW. ¼ sec. 29, T. 12 N., R. 2 W., Monmouth Quadrangle, Illinois; (3) NE. ¼ NE. ¼ sec. 36, T. 9 N., R. 1 W., Avon Quadrangle, Illinois; (4) vicinity of Port Byron, Illinois: *Pecopteris serrulata* (2), *Neuropteris tenuifolia* (early form) (1), *Archaeopteris stricta* (1), *Cannophyllites marginata* (4), *Cannophyllites abbreviata* (4), *C. dawsoni* (1), *C. rectinervis* (2), *C. southwelli* (3, 4), *C. fasciculata* (4), *Alethopteris* sp., *Eremopteris grandis* (2), *Ovopteris communis* (2), *Cardiocarpon* sp. (large) (2, 3), *Trigonocarpus* sp. (2), *Sigillaria rugosa* (2), *Annularia cuspidata* (2), *Lepidodendron aculeatum* (3).

Zone 5.—Zone of *N. tenuifolia*. This zone is characteristic of the major portion of the Kanawha formation and is apparently also characteristic of rocks included in the Lampasas group in the Mid-Continent region.

The following list of species from the shale above the Cannelton coal at Tell City, Indiana, illustrates a typical flora from this zone: *Alethopteris lonchitica*, *A. owenii*, *N. scheuchzeri*, *N. tenuifolia*, *Mariopteris* sp., *Cardiocarpon* sp., *Trigonocarpum* sp., *Lepidodendron* sp., *Lepidophyllum* sp., *Lepidostrobus* sp.

Zone 6.—Zone of *N. rarinervis*. The flora, characterized by *N. rarinervis*, occurs in the lower part of the Allegheny formation in the Appalachian region and in the lower part of the Des Moines group in the Mid-Continent region. It is the highest zone in the suite of Lower Pennsylvanian floras.

The following floras, collected from the roof of the Murphysboro coal at mines near Murphysboro, Illinois, is characteristic of this zone: *Pecopteris vestita*, *Alethopteris serlii*, *Mariopteris occidentalis*, *M. sillimanni*, *Neuropteris ovata*, *N. scheuchzeri*, *N. rarinervis*, *N. clarksoni*, *Linopteris rubella*, *Odontopteris* sp., *Cordaites communis*, *Lepidophyllum oblongifolium*, *Stigmaria ficoides*, *Calamites suckowi*, *Annularia stellata*, *Sphenophyllum emarginatum*.

Zone 7.—Zone of *N. flexuosa* and appearance of abundant *Pecopteris* spp. This zone occurs in the upper part of the Allegheny formation and is characteristic of the upper part of the Des Moines group. It is the lowest zone in the suite of Upper Pennsylvanian floras.

Zones 8 and 9.—Zones of *Lescuropteris* and *Danaeites*, respectively. These two zones are separable in the Appalachian region, where the lower, Zone 9, characterizes the upper Conemaugh and lower Monongahela formations. The upper, Zone 8, is characteristic of the upper part of the Monongahela formation. In the Mid-Continent region the two cannot everywhere be separated and perhaps should be designated as a single unit—the zone of *Odontopteris* spp.

The following list, compiled from collections from the Chanute shale, Kansas City group, illustrates a flora from the lower part of the *Odontopteris* zone:

Mariopteris pluckenetii, *M. cordato-ovata*, *Sphenopteris pinnatifida*, *Aloiopteris winslovii*, *Alethopteris virginiana*, *Alethopteris* sp. cf. *A. virginiana*, *A. grandini*, *Odontopteris reichiana*, *Neuropteris ovata*, *N. plicata*, *N. scheuchzeri*, *Annularia stellata*, *Sphenophyllum oblongifolium*, *Sigillaria camptotaenia*.

FLORAL MODIFICATIONS IN THE ROCKY MOUNTAIN REGION

With this brief sketch of the characteristic floral sequence in the typical Pennsylvanian series, it is now appropriate to examine the floras of rock sequences some distance from Coal Measures deposition. In the southern Rocky Mountains considerable stratigraphic work involving the Pennsylvanian series has been undertaken in recent years. Physical conditions and sedimentation in that area were unlike those of the coal basins. A series of linear positive elements, developed in early Pennsylvanian time, and variable quantities of continental and marine sediments were deposited in the adjacent basins. Certain of the positive areas were sufficiently large to provide considerable tracts of uplands.

Remains of plants that are occasionally found in the clastic portions of the Pennsylvanian sequences afford an opportunity for comparison with floras in the Appalachian and Mid-Continent areas. Since these floras are still in process of investigation, I shall draw from them only a few illustrations bearing on an important point.

In the Weber(?) formation of the Mosquito Range, in central Colorado, a flora belonging to Zone 3, the zone of *Mariopteris pygmaea*, has been described.[3]

At a locality in the lower 100–200 feet of the formation on Evans Peak, the following species were found: *Neuropteris dluhoschi*, *N. heterophylla*, *N. gigantea?*, *Sphenopteris hoeningshausii*, *Sphenopteris* sp. cf. *S. microcarpa*, *Diplothmema cheathami*, *D. patentissima*, *Adiantites rockymountanus*, *Cordaites* sp., *Cordaicarpon* sp., *Trichopitys whitei*, *Dactylophyllum johnsoni*, *Lepidostrobus weberensis*, *Stigmaria verrucosa*, *Calamites* sp., *Asterophyllites longifolius*(?). The dominant species are those of the Coal Measures facies. Fernlike plants and representatives of the Cordaitales are abundant, and remains of Lycopodiales are rare. There are also present, in abundance, the putative conifers *Trichopitys* and *Dactylophyllum*.

In the Upper Pennsylvanian strata of the McCoy formation, north-central Colorado, occur fernlike plants associated with species of the Paleozoic conifer, *Walchia*. The following florule is reported from interval 74, McCoy formation near Yarmony School, Eagle County, Colorado: *Odontopteris mccoyensis*, *Samaropsis hesperius*, *Walchia stricta*, *Walchia* sp., *Walchiastrobus* sp.[4] The flora appears to belong to Zone 7, the zone of *N. flexuosa*, and is therefore, in terms of the typical Pennsylvanian section, equivalent to the flora in the upper part of the Allegheny formation.

In northern New Mexico the Sandia formation and the lower part of the superjacent Madera limestone contain fairly typical Coal Measures floras in which fernlike plants and *Cordaites* sp. are dominant. In the upper part of the Madera limestone floras containing *Walchia* spp. and other conifers occur.

[3] C. B. Read, "A Flora of Pottsville Age from the Mosquito Range, Colo.," *U.S. Geol. Surv. Prof. Paper 185* (1934), pp. 79–96.

[4] C. A. Arnold, "Some Paleozoic Plants from Central Colorado and Their Stratigraphic Significance," *Univ. Mich., Contr. Mus. Paleon.*, Vol. VI (1941), pp. 59–70.

From these notes on the Pennsylvanian floras of the Rocky Mountain region an important conclusion may be drawn. The plant associations, in the older parts of the Pennsylvanian sequences, are similar to those of the Coal Measures but, through the relative rarity of Lycopodiales, suggest a drier habitat than that indicated by the approximately contemporaneous floras in the eastern coal basins. The higher Pennsylvanian floras are striking departures from the plant associations of the same general ages, in the eastern coal basins, as inferred from index forms and independent stratigraphic data. These plants in the western area occur in suites of sediments that were deposited during a period of widespread orogeny in the southern Rocky Mountains. During this period of mountain-building many geologic data indicate a restriction of lowland and an expansion of upland habitats or areas. The floral modifications are in the direction of mesophytic associations.

For the modifications that occur in the Rocky Mountains I propose that the term "Cordilleran flora" be used.

ANTARCTO-CARBONIFEROUS FLORAS

Upper Carboniferous floras of the Southern Hemisphere contrast strikingly with those of the Arcto-Carboniferous province. An outline of the sequence and distribution of the continental Upper Carboniferous deposits is, however, not within the scope of this paper, and it is therefore sufficient to call attention only to the general features of a few representative sequences.[5]

In Argentina the Pagonzo system includes strata of Lower Carboniferous, Upper Carboniferous, Permian, and Triassic ages. Tillites occupy the basal portions of the sequence and appear to be of Lower Carboniferous and early Upper Carboniferous ages. Above are continental deposits of various types, and still higher are some marine beds.

In Brazil the Santa Catherina system is essentially identical with the Pagonzo system of Argentina. It has been divided into the Itararé, Tubarão, and Passa Dois series, which, in turn, have been divided into local units. The Itararé series is a sequence of tillite and interbedded shale and sandstone and is conformably overlain by the Tubarão series, which is the principal coal-bearing unit in the Paraná Basin.

From the lowest part of Stage 1 of the Pagonzo system, floras of probable Chester or late Mississippian age are reported. Succeeding them are associations characterized by an abundance of *Glossopteris* spp., *Gangamopteris* sp., and *Phyllotheca* sp. A similar flora is known from the Itararé tillite in southern Brazil. Itararé series, Paraná Basin, Brazil (1) near Suspiro, Rio Grande do Sul; (2) Teixeira Soares, Paraná: *Gangamopteris obovata* (1), *Glossopteris indica*, *Gl. browniana*, *Gl.* sp., *Phyllotheca* sp., *Brachyphyllum* sp. cf. *B. australe*.

In the upper part of Stage 1 of the Pagonzo system as well as from the Tubarão series in Brazil occur floras that contain, in addition to species of *Glossopteris* and *Phyllotheca*, an abundance of *Pecopteris*, *Zeilleria*, *Annularia*, *Sphenophyllum*, and *Lepidodendron*. Illustrative of the association, the following forms from a single locality are listed: Tubarão series, Cambuhy, Rio das Pedras, Paraná, Brazil;[6] *Pecopteris pedrasiça* (Ar),

[5] C. B. Read, "Plantas fósseis do Neo-Paleozóico do Paraná e Santa Catarina," *Brasil, Min. Agr., Dept. Nac. Prod. Min., Div. Geol. Min., Mono. 12* (1941), pp. 1–102.

[6] In the list forms indicated by "I" are known by the writer to occur in the Itararé series. Forms indicated by "An" are considered typical members

P. cambuhyensis (Ar), *P. paranaensis* (Ar), *Zeilleria oliveirai* (Ar), *Glossopteris indica* (I, An), *Glossopteris* sp. (I, An), *Glossopteris* sp. cf. *ampla* (I, An), *Sphenophyllum oblongifolium* (Ar), *Annularia? americana* (Ar), *Phyllotheca australis* (I, An), *Phyllotheca* sp. (I, An), *Brachyphyllum* sp. cf. *B. australe* (I), *Buriadia* sp. (An), *Lepidostrobus* sp. (Ar, An), *Lepidodendron pedroanum* (Ar, An).

A preliminary conclusion is that such floras as those from the upper part of Stage 1, Pagonzo system, and from the Tubarão series, Santa Catherina system, are correlative with the upper part of the Pennsylvanian series of North America. This conclusion is based on the general similarity of the types common in the floras. It is further concluded that the tillite sequence which contains a zone of abundant *Glossopteris*, *Gangamopteris*, and *Phyllotheca* is also of Pennsylvanian age.[7]

The presence of a flora so dissimilar to those known in the Arcto-Carboniferous province, succeeded by others having many similarities to Arcto-Carboniferous floras, is perhaps explained by reference to the rock sequences. The floras of the Itararé series and its equivalents are in shale interbedded with tillite. The environment, although probably not then glacial in the area inhabited by the plant associations, was probably more rigorous than is generally characteristic of Carboniferous floras. It is apparent from a consideration of paleogeography that the Antarcto-Carboniferous area in question was separated from Arcto-Carboniferous land areas; therefore, opportunities for development of dissimilar floras were present. Following maximum glaciation in the Pennsylvanian of South America, climatic amelioration occurred, as indicated by the presence of coal in the superjacent rocks. During, or prior to, that period of time, members of the Arcto-Carboniferous floras presumably migrated into the Southern Hemisphere and became common in the known plant associations.

SUMMARY

Two major floral provinces have been recognized as existing during the Pennsylvanian epoch: (1) the northern or Arcto-Carboniferous warm, rain forest province of coal floras and (2) the southern or Antarcto-Carboniferous cool, rain forest province. In the Arcto-Carboniferous province the dominant associations preserved are Coal Measures floras. The eastern and Mid-Continent regions appear to have been a rather continuous lowland, across which shallow seas intermittently advanced and retreated, and physical conditions must have been generally similar everywhere in the basin. Thus widespread migration of plant associations in negligible time is probable, and the floral zones may be inferred to have chronologic value.

In portions of the Rocky Mountain region there were restricted lowlands and basins of deposition adjacent to rapidly rising geanticlines. Modifications of the lowland floras are found there, these modifications being due to the inclusion of mesophytic forms. The presence of such types may be explained by the existence of upland habitats, with mesophytic floras growing upon them, adjacent to sites of deposition; it may also be interpreted in terms of regional climatic variations.

To the sequence of rain forest lowland or Coal Measures floras the term "Arcto-

of the Antarcto-Carboniferous flora; by "Ar" are those related closely to Arcto-Carboniferous types. Several of these are conspecific with, or closely related to, forms that occur in the Arcto-Carboniferous floras of the Pennsylvanian series.

[7] Read, ftn. 5.

Carboniferous" has been applied because of the widespread occurrence of this suite of associations in the Pennsylvanian or Upper Carboniferous rocks of the Northern Hemisphere. For the modifications that occur in the southern Rocky Mountains the term "Cordilleran flora" is recommended.

The Antarcto-Carboniferous floras were, during earlier Pennsylvanian time, dominantly cool, rain forest types that developed independently or semi-independently of the Arcto-Carboniferous floras. During later Pennsylvanian time there was sufficient migration of Arcto-Carboniferous floras southward to modify the earlier existing floras. This occurred at a time of climatic amelioration following glaciation.

More important than the discussion of these wanderings of the Pennsylvanian floras are the principles that have been suggested. Geographical distribution is a frequently neglected factor in studies of Paleozoic organisms. Furthermore, differences between floras on the two margins of a single major depositional area are sufficiently great to require the utmost care in establishing even approximate correlations. Such care consists of weighing all available evidence, organic and stratigraphic, and of study of entire associations rather than of certain selected biologic groups.

It is perhaps inadvisable to draw any generalization from these observations. Depending upon one's point of view, it might be concluded that land floras show so many variations that regional correlations based on them are of doubtful value. Equally possible is the conclusion that problems of geographical distribution and variation have not been sufficiently considered by paleontologists.

PLANT MICROFOSSILS IN CORRELATION OF COAL BEDS[1]

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ABSTRACT

An intensive study of fossil microspores, isospores, and prepollens isolated from Illinois coal beds has proved that numerous genera and species are present. The correlation of coal beds is possible because the geological range of many genera and species is proved, by this investigation, to be quite restricted.

Preliminary experiments with specific interregional correlations have demonstrated a wide lateral distribution of plant microfossils necessary for the correlations of this type.

INTRODUCTION

The purpose of this paper is to show that plant microfossils may be used effectively in the correlation of spore-bearing strata. The extent to which they may be used interregionally is of utmost importance if rapid progress is to be made in stratigraphic studies. These preliminary results will therefore help to evaluate the role which plant microfossils can play in interregional correlations. A final report will include descriptions of the many new genera and species of plant microfossils isolated from Illinois coal beds and the methods of isolating them for study.

Although plant spores were observed in coal more than a hundred years ago, it is only recently that these microscopic remains have been used by European geologists in solving stratigraphic problems. Their use is practically unknown in this country. This is readily understood by paleobotanists familiar with the numerous taxonomic problems which they present. A recent publication[2] has considerably lessened the taxonomic problems, so that investigators may now place plant spores in their appropriate genera. The present investigation is the first comprehensive survey of isolated microspores, isospores, and prepollens of Paleozoic strata in this country. J. M. Schopf,[3] using megaspores, indicated their possible use in correlating coal beds. The first correlation studies of two Ohio coal beds based on isolated small spores proved successful.[4]

DISTRIBUTION OF SMALL-SPORE GENERA IN ILLINOIS COAL BEDS

THE CASEYVILLE GROUP

The oldest Pennsylvanian strata in Illinois contain ten established genera and at least three new genera. *Densosporites* and *Lycospora* are the dominant genera, and thus in all probability the coals of this group were largely derived from the parent-plants of these two genera. The parent-plant of *Densosporites* is unknown, but certainly the spores of *Lycospora* are those of the arborescent lepidodendrons.

The Wayside, Battery Rock, and

[1] Published with the permission of the chief, Illinois Geological Survey, Urbana, Illinois.

[2] J. M. Schopf, L. R. Wilson, and Ray Bentall, "An Annotated Synopsis of Paleozoic Fossil Spores and the Definition of Generic Groups," *Ill. Geol. Surv. Rept. Investigations No. 91* (1944), pp. 1–66.

[3] "Spores Characteristic of Illinois Coal No. 6," *Trans. Ill. Acad. Sci.*, Vol. XXVIII, No. 2 (1936), pp. 173–76; "Spores from Herrin (No. 6) Coal Bed in Illinois," *Ill. Geol. Surv. Rept. Investigations No. 50* (1938), pp. 1–73.

[4] R. M. Kosanke, "The Characteristic Plant Microfossils of the Pittsburgh and Pomeroy Coals of Ohio," *Amer. Mid. Nat.*, Vol. I (1943), pp. 119–32.

Reynoldsburg coal beds from southern Illinois are considered to be of Caseyville age, which is in agreement with J. M. Weller, L. G. Henbest, and C. O. Dunbar.[5] The coal beds of the Caseyville group appear to lack species of the genus *Cirratriradites*, and the genus *Laevigato-sporites* may have originated in Reynoldsburg time. At least, *Laevigato-sporites* is represented in the Caseyville by only a few specimens, which is surprising, because spores of this type are known to have an almost continuous range from the Pennsylvanian to the present day.[6] If *Laevigato-sporites* originated in Reynoldsburg time, there must have been a major floral change. Such a floral change supports the correlation[7] of the Babylon coal bed as post-Caseyville, because the genus is abundantly represented in the Babylon coal bed, as is the genus *Cirratriradites*. The Babylon coal bed was formerly considered to be Caseyville.[8]

THE TRADEWATER GROUP

The Tradewater group contains species of fourteen established genera and at least two new genera which are present in the Caseyville group. The flora was more diversified, and the dominant genus throughout most of this period was *Laevigato-sporites*, which was scantily represented earlier. Only *Lycospora* replaces *Laevigato-sporites* as the dominant floral element in one coal bed. The subdominant genera during part of this time are *Granulati-sporites*, *Cirratriradites*, *Triquitrites*, and *Lycospora*. In most cases each of the momentary periods of subdominance indicates the origin of new species, which serve as guide fossils. Also *Alati-sporites*, *Cirratriradites*, *Florinites*, and possibly *Reinschospora* appear for the first time during the Pennsylvanian period in Illinois.

The Willis and Tartar, Pope Creek and Delwood, Seville and Rock Island coal beds are to be correlated on the basis of plant microfossils, which is in agreement with C. L. Cooper.[9] The Curlew coal bed is thought to be younger than the Rock Island and Seville coal beds,[10] and this view is held by the author. The Bald Hill coal bed, originally described by G. H. Cady,[11] is younger than the Curlew coal bed. The Davis and Wiley are correlated and likewise the DeKoven and Greenbush; this is in agreement with Wanless' suggested correlations.[12]

THE CARBONDALE GROUP

The coal beds of the Carbondale group present fewer correlation problems than any other group in Illinois. One new genus is of utmost importance in correlating the No. 2 coal bed and is known to occur only in the lower half of the bed. This coal bed has species of *Lycospora* as the dominant microfloral element; and a species of *Laevigato-sporites*, with its origin in the DeKoven coal bed,

[5] "Pennsylvanian Fusulinidae of Illinois," *Ill. Geol. Surv. Bull. No. 67* (1942), pp. 9–34.

[6] L. R. Wilson and R. M. Webster, "Plant Microfossils from a Fort Union Coal of Montana," *Amer. Jour. Bot.*, Vol. XXXIII, No. 4 (April, 1946), pp. 271–78; Olof H. Selling, "Spores of Hawaiian Pteridophytes. I." *Spec. Pub. 37, Bishop Museum, Honolulu, Hawaii* (Göteborg, Sweden, 1946), pp. 1–87.

[7] R. C. Moore, H. R. Wanless, J. M. Weller, *et al.*, "Correlation of Pennsylvanian Formations of North America," *Bull. Geol. Soc. Amer.*, Vol. LV (1944), pp. 657–706.

[8] Weller, Henbest, and Dunbar, ftn. 5; Harold R. Wanless, "Pennsylvanian Correlation in the Eastern Interior and Appalachian Coal Fields," *Geol. Soc. Amer. Spec. Paper 17* (1939), pp. 85 and 108.

[9] "Pennsylvanian Ostracodes of Illinois," *Ill. Geol. Surv. Bull. 70* (1946), p. 16.

[10] C. O. Dunbar and L. G. Henbest, "Pennsylvanian Fusulinidae of Illinois," *Ill. Geol. Surv. Bull. 67* (1942), pp. 20–21.

[11] "The Areal Geology of Saline County," *Trans. Ill. Acad. Sci.*, Vol. XIX (1926), pp. 259–60.

[12] Ftn. 8 (1939).

becomes a prominent member of the flora. Number 5 coal bed is readily distinguished by an abundance of *Laevigatosporites*, fewer *Lycospora*, and the presence of several species of *Alati-sporites* and *Cirratriradites*. Number 5-A coal bed is distinct from both No. 5 and No. 6 coal beds. It does not correlate with the Grape Creek coal bed, which is in contrast to the correlations of J. M. Weller and H. R. Wanless[13] and of Wanless.[14] The correlation of the Grape Creek coal bed with No. 6 coal bed by F. H. Bradley[15] is upheld by spore studies in which no generic or specific differences have been noted. The Grape Creek coal bed has been variously correlated as No. 5 coal bed[16] and No. 5-A coal bed.[17] The large spores of No. 6 coal bed, megaspores, are known through the works of J. M. Schopf.[18]

THE McLEANSBORO GROUP

In the McLeansboro group there are more stratigraphic problems than in any other group in Illinois. Much additional work must be done before the problems can be solved. However, there are fourteen established genera known to occur and at least one new genus is present. Changes in the distribution and abundance of genera are most encouraging. Danville No. 7 coal bed is definitely younger than the Jamestown or Bankston Fork coal beds. It may be identical with the Cutler coal bed or the first Cutler rider coal bed. There exists a major floral break just below No. 8 coal, indicated by the fact that no species of *Lycospora* has as yet been observed in No. 8 coal. This is the approximate position of the boundary of the Des Moines and Missouri series of the Mid-Continent, according to Cooper.[19] Coal samples from type areas of the Shoal Creek and Carlinville show that they do not correlate and that they are separate and distinct strata. Further, they show that the Carlinville is probably older than the Shoal Creek. The presence of the genus *Pityosporites* in the Shoal Creek and younger beds strongly suggests the presence of an upland conifer flora.

SUMMARY OF SMALL-SPORE INVESTIGATIONS IN ILLINOIS

These remarks are based on more than six hundred macerations from forty-seven counties in the state. A study of Figure 1 readily reveals that, through a knowledge of the genera and their abundance, one can place an unknown coal bed in one of the four groups of the Pennsylvanian system, and in many cases even identify it specifically. A knowledge of the numerous species and their abundance readily permits the correlation of Pennsylvanian spore-bearing strata in Illinois.

PRELIMINARY INTERREGIONAL CORRELATIONS

After a comprehensive survey of the spore content of Illinois coal beds had been made, it was important to investigate the possibility of using plant remains isolated from spore-bearing strata

[13] "Correlation of Minable Coals of Illinois, Indiana, and Western Kentucky," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XXIII, No. 9 (1939), pp. 1374–92; reprinted in *Ill. Geol. Surv. Circ. No. 48* (1939).

[14] Pp. 84 and 110 of ftn. 8.

[15] "Geology of Vermilion County," *Ill. Geol. Surv.*, Vol. IV (1870), pp. 241–65.

[16] H. R. Wanless, quoted in G. H. Cady, "Classification and Selection of Illinois Coals," *Ill. Geol. Surv. Bull. 62* (1935), p. 39.

[17] Weller and Wanless, ftn. 13; and Wanless, ftn. 8.

[18] Ftn. 3.

[19] Ftn. 9.

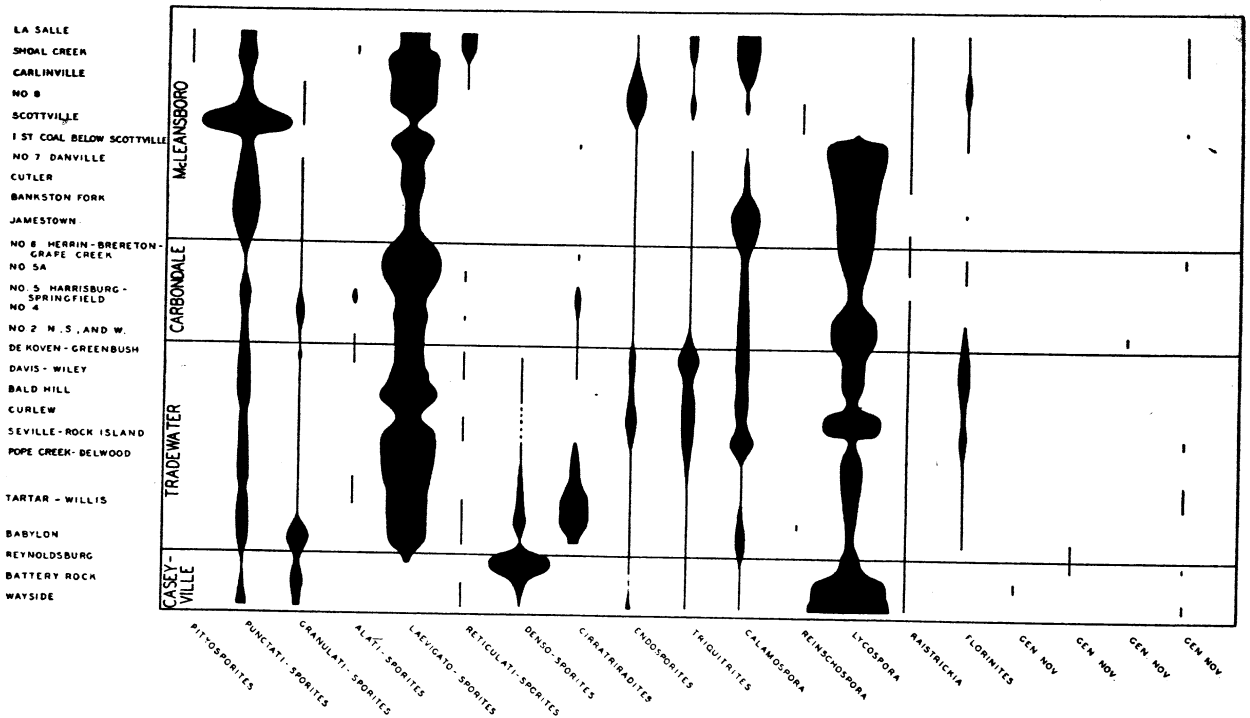

FIG. 1.—Distribution and relative abundance of the small-spore genera in a number of Illinois coal beds. The coal beds are listed to the left and the genera at the base. The genus *Punctati-sporites* for the Wayside coal bed represents 10 per cent, and this may be used as a scale.

for interregional correlations. Coal samples from Ohio, Iowa, and Missouri were prepared and examined. The results are encouraging and may be summarized as follows: Characteristic plant spores are present in the coal beds investigated, although their abundance is thought to be at variance with comparable horizons in Illinois. This may be due to ecological conditions. The preliminary tests suggested possible correlations with certain Illinois beds, but, because of limited samples, specific correlations should not be made at this time. Microfloral zones, known from Illinois coal beds, may be used to correlate interregionally.

CONCLUSIONS

Correlation of Pennsylvanian spore-bearing strata is possible by the use of microspores, isospores, and prepollens when prepared by the maceration method. This type of research gives promise of being a useful tool to the geologist and is of importance to the paleobotanist in determining the range and abundance of fossil plants. Numerous investigations of this kind should be carried out in critical areas throughout the country.

SPORE FLORAS OF THE PENNSYLVANIAN OF WEST VIRGINIA AND KENTUCKY

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ABSTRACT

Each of nearly one hundred coal seams of Pennsylvanian age has been sampled at several localities in West Virginia and adjacent states. Maceration residues of these coal samples show a differentiation of the microfossil flora between the seams, which may represent evolutionary or/and ecological changes. Preliminary studies have been made on the megaspores which permit some advance conclusions. Sixty species and varieties of the megaspore flora of twenty selected seams are illustrated and discussed. Fifteen types of microspores from three seams are given from among fifty-eight recognized in all the coals studied. The principal floral changes appear to occur above the Lower War Eagle and Buffalo Creek coals and below the Uniontown coal. Less important changes are noted above the Matewan, Cedar Grove, and Lower Kittanning coals. Some evidence for interregional correlation based on similar geologic occurrence of certain spores is given. Several problems of the morphology and identification of spores are presented in the illustrated discussion of spore characteristics.

INTRODUCTION

The investigation of the microfossils of the Pennsylvanian coal seams of a portion of the Appalachian coal field has been under way since August, 1940. The principal study has been devoted to the coals of the Kanawha (Upper Pottsville) and Lower Allegheny series of southern West Virginia and eastern Kentucky. Since circumstances during these years have hindered the work, this paper must be understood as a report of progress. The researches in hand may be divided into several projects, each of which will be published as it is completed. The objectives were stated before an earlier meeting of geologists[1] and need not be reviewed here. The project which deals with the identification and classification of the microfossil assemblages is in itself a complex problem. The procedure and progress of this study to date are briefly presented.

COLLECTIONS

Early field work was limited to collecting samples of coal from mines in economically important coal seams, whose advertised identifications were presumed to be correct. Before work had proceeded far it was found that these identifications were too unreliable for a detailed study. All subsequent collections, over 750 samples from the area shown in Figure 1, were made from horizons which could be determined with reasonable accuracy by established methods of correlation. The geographic distribution of the samples thus far examined is also shown in Figure 1. Locality numbers given there are not those on the filed samples but, instead, are arbitrary designations for convenience in this report. At Localities Nos. 1–23, 28, 31–35, and 39, individual collections from particular coal seams were made. Numbers 24–27, 30, 36–38, and 40 refer to localities where thick series of exposed strata include a number of coal seams. Locality No. 29 includes three complete sections taken from three closely spaced drill cores. Several hundred additional collections awaiting further examination are not indicated on this figure.

Figure 2 shows the geologic distribution of some of the coals sampled and

[1] American Association for the Advancement of Science, Geology Section, Cleveland, 1944.

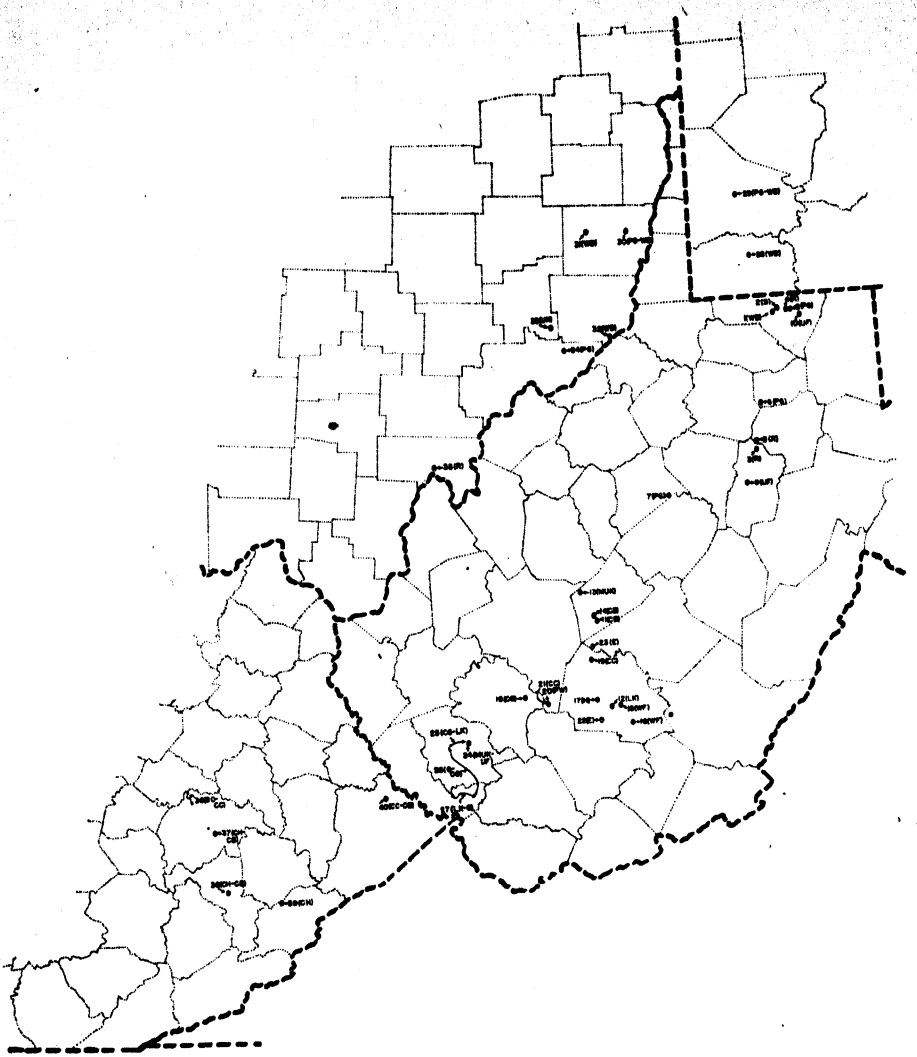


FIG. 1.—County outline map of most of West Virginia and adjacent areas of Kentucky, Ohio, and Pennsylvania, showing localities (Nos. 1–40) from which coal samples were collected for study of the microfossil floras. The letters in parentheses after each number indicate which of the coals listed on the chart in Fig. 2 have been collected at each locality. Abbreviations of names of coal seams are as follows:

Waynesburg	*WB*
Uniontown	*UN*
Sewickley	*S*
Redstone	*R*
Pittsburgh	*PG*
Upper Freeport	*UF*
Upper or Middle Kittanning	*MUK*
No. 5 Block (L. Kittanning)	*LK*
Coalburg	*CB*
Buffalo Creek	*BC*
Winifrede	*WF*
Chilton	*CH*
Hernshaw	*H*
Cedar Grove	*CG*
Campbell Creek	*CC*
Powellton	*PW*
Eagle	*E*
Cedar	*C*
L. War Eagle	*LWG*
Gilbert	*G*

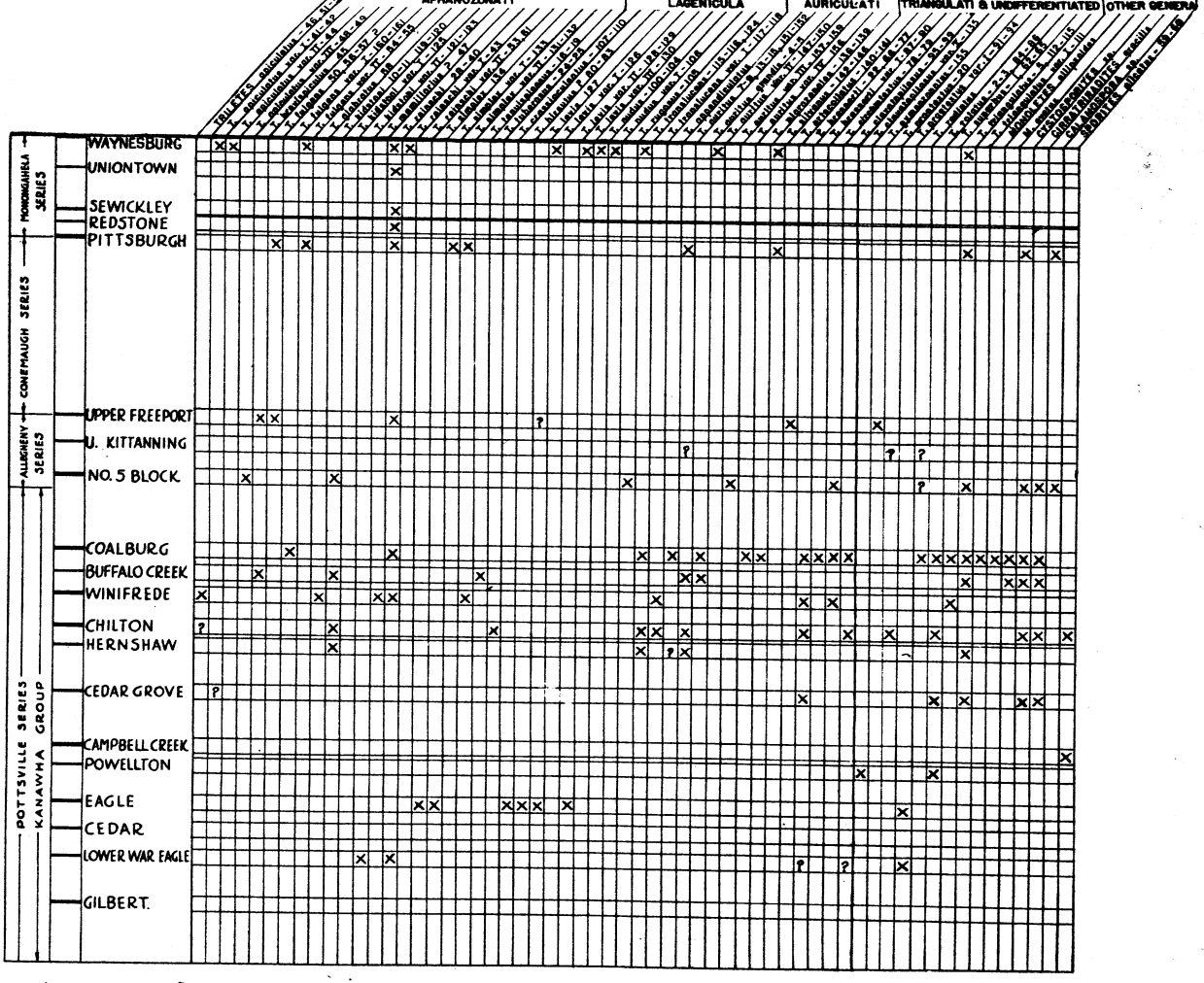


FIG. 2.—Chart placing twenty of the coal seams shown on map (Fig. 1) in relative position in the geologic column. The occurrence and distribution of megaspores of each of 54 species or varieties of *Triletes* and 6 species of 5 other genera are indicated by an × or by a "?" where specific identity is uncertain. Many other unidentified spores have been found which are not recorded here.

examined. In the excellent section given by David B. Reger[2] for the Pennsylvanian of West Virginia, 91 cycles of deposition are shown. These include a total of 102 named coal horizons. Most of these cyclothems contain a principal coal seam, and many, especially in the Kanawha group, include an additional coal, which may be locally important. Altogether, about 60 seams in West Virginia have been mined or might be considered to be of economic value. It will be noted in Figure 2 that the occurrence of microfossils is given for only 20 of these. Samples have been examined from all economic coal seams of the Pennsylvanian system shown by Reger, except the Little Clarksburg and the Elk Lick coals of the Conemaugh and the coals of the Pocahontas group. For reasons which are not clear, the microfossils separable from the residues of the coals below those of Kanawha age are not distinct or numerous in the samples examined. It is likely that the increase in rank of coal has been accompanied by volatilization of most of the waxy materials, and it is also possible that the early Pennsylvanian floras and/or their environmental conditions may have been sufficiently different from those which followed to explain in part the observed paucity of microfossils.

In several places it has been possible to obtain samples from a series of consecutively younger coal horizons, which are exposed almost continuously along recent road cuts and railroads. In Logan and Mingo counties in southwestern West Virginia, a relatively continuous outcrop of mid-Pennsylvanian strata is exposed for nearly 40 miles (Localities Nos. 24–27, Fig. 1). From Kanawha beds at the eastern base of Blair Mountain, a sequence can be traced up to the Upper Allegheny strata and back down to mid-Kanawha in the traverse of U.S. Highway 119 (and the fire-tower road) over the mountain to the Guyandot River at Stollings. From there, owing to the regional dip, consecutively older strata nearly to the base of the Kanawha group are exposed in almost continuous sequence upriver to Gilbert. From Gilbert the section continues back up through all the Kanawha into the Lower Allegheny strata at the top of Horsepen Mountain. In the complete traverse all coal horizons from the Gilbert to the No. 5 Block (Lower Kittanning) are exposed at two or more places, and, in addition, coals up to the Lower Freeport are exposed along the fire-tower road on Blair Mountain. All 37 coals or coaly horizons exposed have been sampled at each locality, and some of the results are recorded here.

Nearly comparable series of exposures in Breathitt and Pike counties, Kentucky (Localities Nos. 36 and 40, Fig. 1), have been studied, and the results of preliminary examination of coal residues are included here.

A relatively continuous section of Monongahela strata from Belmont County, Ohio (Locality No. 30, Fig. 1), recently published by George W. White,[3] and some even more complete sections taken from drill cores (Locality No. 29, previously mentioned) in Washington County, Pennsylvania, show all named coal zones and some additional coaly horizons of the Monongahela and all the Washington formation coals of the Permian strata above. The examination of the macerated residues from these coals

[2] "Pennsylvanian Cycles in West Virginia," *Ill. Geol. Surv. Bull. 60* (1931), pp. 220–31, Figs. 51–52.

[3] "Upper Pennsylvanian and Lower Permian Rock Section at Blaine Hill, Belmont County, Ohio," *Ohio Jour. Sci.*, Vol. XLV (1945), pp. 173–79.

is in progress, and some results are available for this report.

NATURE OF SAMPLES

The samples collected from the various coal horizons differ in nature. Most good exposures of coal seams were sampled by benches, or, where definitive boundaries are absent, samples were taken at regular intervals through the seam. For instance, the Pittsburgh coal seam, in regions of good development of its partings and overlying "A," or Rider and Rooster coals, has been divided to yield as many as 17 samples at one exposure, whereas in the poorly developed marginal areas of the field, such as at Locality No. 34 (Fig. 1) in Monroe County, Ohio, only a channel sample was deemed to be useful. Occasionally, a channel sample of a thick seam was taken, especially where exposure was poor and either the top or the bottom of the bed was concealed. In many instances it was possible to obtain only a sample of the "blossom" of a coal seam, i.e., a mixture of weathered coal fragments and soil mantle. These fragments appear to be satisfactorily representative of the coal from which they are derived and are much easier to prepare for study than are unweathered coal samples.

Many of the samples obtained from the long series of exposed strata discussed above represent thin streaks of attrital coal or coaly shales. Coaly streaks, which are determined to be merely single vitrain bands, a common occurrence in roof shales and sandstones, were not sampled, for it is assumed, and in many cases it is demonstrable, that they usually represent individual coalified fragments of allochthonous (drift) vegetal matter, such as stems and leaves. The thin attrital coals and coaly shales were sampled, for they are often representative of a coal horizon which is better developed elsewhere. Such horizons may often be identified in the field by their positions relative to other recognizable key beds. Occasionally, such layers may be discovered by special examination in thick series of shales which are without distinctive breaks, thus bridging facies changes and apparent depositional unconformities, where several beds, including a coal seam, appear to be missing.

PREPARATION AND STUDY OF SAMPLES

A portion of each sample was subjected to a maceration treatment for freeing the microfossils from the other coal constituents. Such techniques have been described by various workers, so it will suffice to say here that satisfactory results are obtained by using any of several variations of the Schulze[4] or Reinsch[5] methods, depending on the nature of the sample. Weathered coals break down more quickly and more satisfactorily than unweathered samples of the seams. Splint coals are more resistant to the oxidation process than are most other coal types. A complete review of the maceration techniques employed on all ranks and types of coal and a statistical summary of results is in preparation.

Each residue was separated into two sizes by being passed through a screen (Tyler Standard No. 65, opening *ca.* 208 μ). The coarser residues contained a variety of plant fragments—large spores, cuticles of leaves and stems, bits of wood (fusain), fragments of the vascular ele-

[4] Franz Schulze, "Über das Vorkommen wohlerhaltener Zellulose in Braunkohle und Steinkohle," *Ber. k. Akad. Wiss. Berlin* (1855), pp. 676–78.

[5] P. F. Reinsch, *Micro-palaeophytologia formationis carboniferae*, Vol. I: *Continens Trileteas et Stelideas* (Erlangen: Theo. Krische, 1884).

ments, etc.[6] These residues were examined under Greenough type binoculars, and recognizable plant remains were lifted out with special needles or brushes. The selected microfossils were then stored in vials of glycerine-alcohol; mounted dry on regular, black-cardboard microfossil slides; or, if translucent, mounted in thick balsam or glycerine jelly on glass slides and covered with cover slips. The fine residues were transferred to vials of glycerine-alcohol, and, from these, small portions were stained and mounted in Canada balsam or Diaphane on regular slides.

In identification of the microfossils it is necessary to examine a rather large number of specimens of the same type in order to understand possible variations and differences of appearance depending on the nature of preservation. Some specimens may be flattened normally in one direction, but occasionally they are compressed in a different plane and thus are of considerably different appearance. Specimens may be broken, partially or completely stripped of characteristic ornamentation, or distorted by the effect of adjacent sand grains or hard plant fragments. Often it is necessary to disregard some of the problematical forms as undeterminable. Some microfossils are consistently well preserved and complete, others as consistently poorly preserved. Such conditions are in line with those to be expected, according to observations made on recent deposits, such as bog and lake sediments.

IDENTIFICATION AND CLASSIFICATION OF MICROFOSSILS

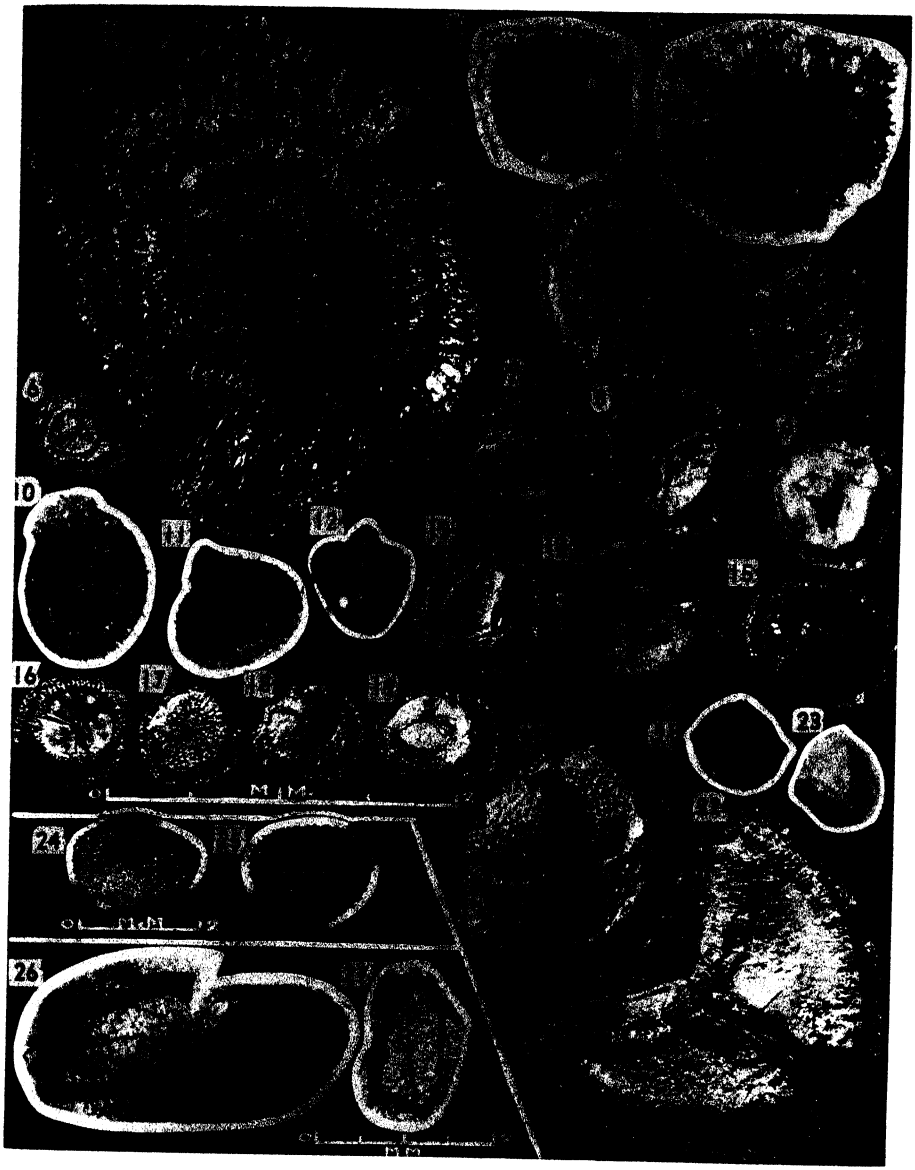
The principal identifiable plant parts preserved in coal are cutinized tissues, pollen grains, spores, and altered wood tissues. During maceration processes these are freed from the complex black humic substances, mixtures of partially decomposed plant remains, and mineral matter. The cuticular remains, i.e., cutinized epidermal tissues and continuous waxy layers (cuticles), are presumed to be of value for determination of the types of plants represented and may be shown to be of index value for coal seams. Isolation of large numbers of these from various coal seams has been carried on, but no identification has yet been attempted in the present study.

Pollen grains and spores are not homologous structures. Therefore, it is desirable to be specific as to which is under consideration. A strict botanical definition here would probably be confusing; hence it may be said, with reservations, that pollen grains are microscopic plants composed of several cells enclosed in a special covering, which develop from spores that have not been shed from the male portion of a flower (stamen, microsporophyll). The pollen grain itself is shed and transferred to the female flower structure (pistil, megasporophyll) by gravity, wind, or insects. Spores, as they are shed from ferns, mosses, and other plants of lower evolutionary rank than the seed plants, are unicellular, microscopic bodies enclosed by a special covering of somewhat similar nature to that found on pollen grains. Spores of these lower plants are shed and carried about by air currents (or distributed by gravity in the case of a few very large types). If they eventually come to rest on a suitable medium, they may germinate and grow into the sex-bearing (gametophyte) plant of the life-cycle. James M. Schopf[7] gives a good discussion of the

[6] Louis C. McCabe, "Some Plant Structures of Coal," *Trans. Ill. Acad. Sci.*, Vol. XXIV (1931), pp. 321–26, Pls. I–II.

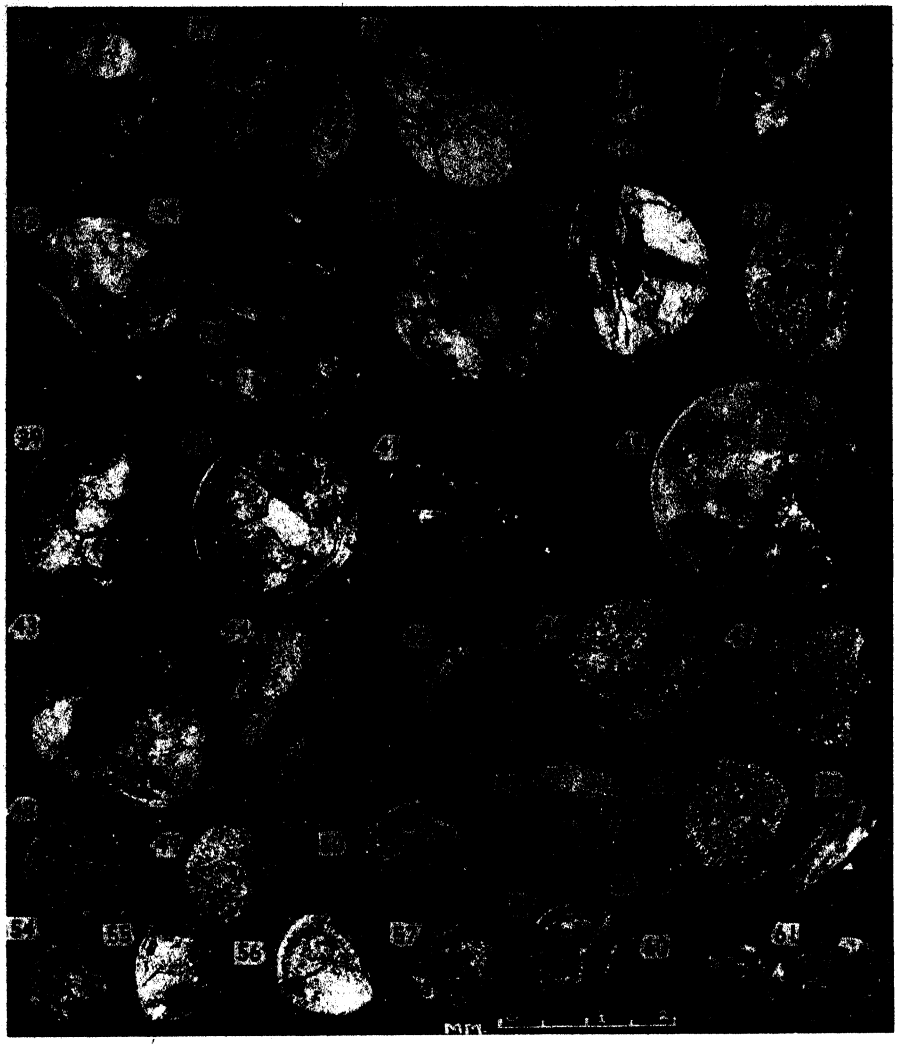
[7] "Spores from the Herrin (No. 6) Coal Bed in Illinois," *Ill. Geol. Surv., Rept. Investigations No. 50* (1938), pp. 10–15.

PLATE I



FIGS. 1–27.—Figs. 1–25, assortment of megaspores of the genus *Triletes* assembled to illustrate range of size, form, and ornamentation. 1, *T. superbus;* 2–3, *T. rotatus;* 4–5, *T. auritus grandis;* 6, *T. triangulatus;* 7–9, *T. auritus;* 10–11, *T. kidstoni;* 12, *T. levis*(?); 13–15, *T. auritus;* 16–19, *T. tenuispinosus;* 20, *T. praetextus;* 21, unidentified species; 22, *T. brasserti;* 23, undetermined species; 24–25, *T. tuberosus.* Figs. 26–27, unidentified sporangia.

PLATE II

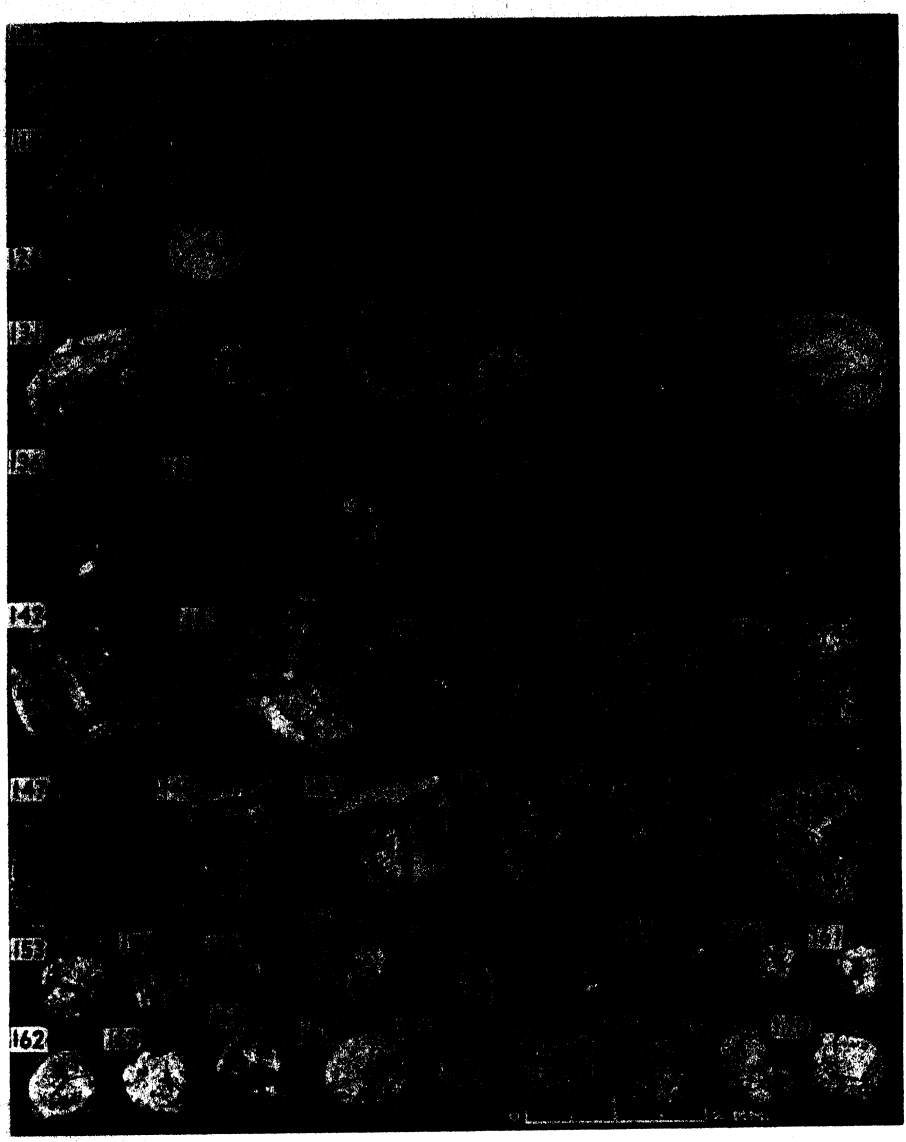


FIGS. 28–61.—Figs. 28–58, 61, genus *Triletes*, section *Aphanozonati;* Figs. 59–60 spores of uncertain affinity. 28–40, *T. reinschi;* 41–42, *T. apiculatus* var. I; 43, *T. reinschi* var. I; 44, *T. apiculatus* var. II; 45, *T. brevispiculus;* 46, *T. apiculatus;* 47, *T. mamillarius*(?); 48–49, *T. apiculatus* var. III; 50, *T. fulgens;* 51–52, *T. apiculatus;* 53, *T. reinschi* var. II; 54–55, *T. fulgens* var. II; 56–57, *T. fulgens*(?); 58, *T. glabratus;* 59–60, *Sporites plicatus;* 61, *Triletes reinschi* var. II.

PLATE III

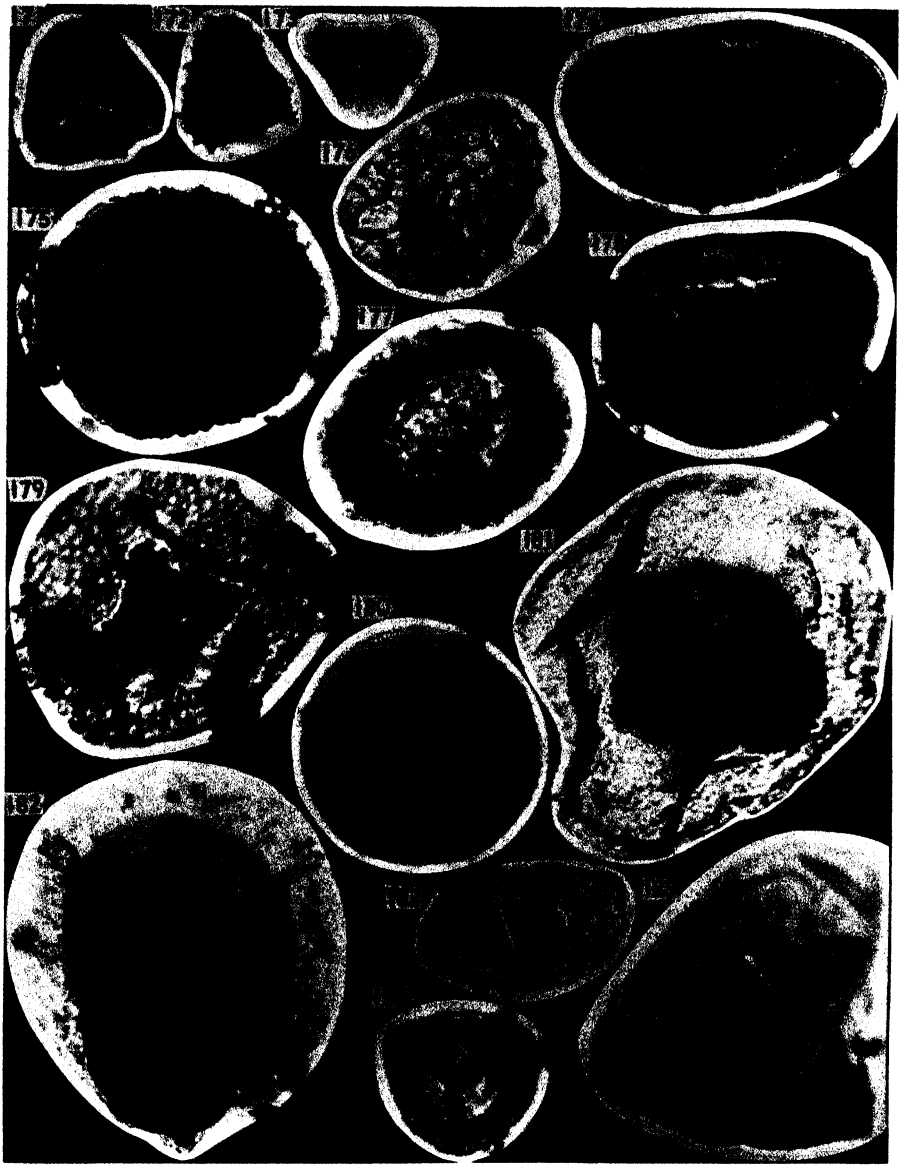
FIGS. 62–106.—Figs. 62–79, 84–99, genus *Triletes*, section *Triangulati* and undifferentiated spores; Figs. 80–83, 100–106, section *Lagenicula*. 62–65, *T. superbus;* 66–77, *T. brasserti;* 78–79, *T. circumtextus;* 80–83, *T. hirsutus*(?); 84–86, *T. rotatus;* 87–90, *T. brasserti* var. I; 91–94, *T. praetextus* var. I; 95–99, *T. clavatopilosus;* 100–104, *T. nudus;* 105, *T. rugosus;* 106, *T. nudus* var. I.

PLATE IV



FIGS. 107–70.—Figs. 107–10, 115–18, 124, 126–30, genus *Triletes*, section *Lagenicula;* Figs. 136–39, 142–52, 156–59, section *Auriculati;* Figs. 111–14, 135, 140–41, 155, section *Triangulati;* Figs. 119–23, 125, 131–34, 160–61, section *Aphanozonati;* Figs. 153–54, 162–70, undetermined species. 107–10, *T. crassiaculeatus;* 111, *T. triangulatus* var. I; 112–14, *T. triangulatus;* 115–116, *T. translucens;* 117–18, *T. translucens* var. I; 119–20, *T. kidstoni;* 121–23, *T. kidstoni* var. I; 124, *T. translucens;* 125, *T. kidstoni* var. I (occurs in Lower War Eagle Coal); 126, *T. levis* var. I; 127, *T. levis;* 128–29, *T. levis* var. II; 130, *T. levis* var. III; 131–32, *T. simplex* var. II; 133, *T. simplex* var. I; 134, *T. simplex;* 135. *T. clavallopilosus* var. I; 136–39, *T. nigrozonales;* 140–41, *T. artecollatus;* 142–46, *T. silvanus;* 147–50, *T. auritus* var. II; 151–52, *T. auritus;* 153–54, undetermined species; 155, *T. gymnozonatus;* 156, T. *auritus* var. IV; 157–59, *T. auritus* var. III; 160–61, *T. fulgens* var. I; 162–70, undetermined species.

PLATE V



FIGS. 171–85.—Representative microspores from the Powellton, Campbell Creek and Alma coals selected to illustrate a number of genera from the Appalachian Basin. 171, *Punctati-sporites* sp., 33 × 36 μ; 172, *Granulati-sporites granulatus*, 20 × 24 μ; 173, *Granulati-sporites* sp., 17 × 20 μ; 174, *Laevigato-sporites* sp., 43 × 75 μ; 175, *Granulati-sporites* sp., 80 × 90 μ; 176, *Reticulati-sporites* sp., 43 × 50 μ; 177, *Denso-sporites* sp., 39 × 50 μ; 178, *Laevigato-sporites* sp., 46 × 60 μ; 179, *Denso-sporites* sp., 53 × 58 μ; 180, unidentified *Sporites*, 53 × 58 μ; 181, *Endosporites ornatus*(?), 115 × 125 μ; 182, *Cirratriradites* sp., 110 × 115 μ; 183, *Endosporites* sp., 30 × 50 μ; 184, *Lycospora confinis*, 37 × 42 μ; 185, *Calamospora* sp., 105 × 115 μ.

morphology and terminology of spores and their equivalents. It is often not possible to distinguish whether one has spores or pollen grains under observation, for usually nothing more than the multiple coat or covering is preserved. The presence of cell contents or structures is required for certain determination.

The spores of a plant may be all alike in appearance, a condition called "homospory" or "isospory," or they may be divisible into two groups originally distinguished on the basis of size, a "heterosporous" condition. The larger spores are called "megaspores" and are more often the female spores. "Microspores," the smaller type in the heterosporous condition, are usually male. A consideration of the fact that microspores have been shown to be larger than their corresponding megaspores in some instances—an important point in botanical studies—would seriously confuse the points being drawn here and so will not be discussed. The generalizations made are adequate to introduce required terminology. The sexes cannot be conclusively proved for fossils, and, further, the microspores are indistinguishable from "isospores" in many cases. The study of the microspores has not progressed far enough to give any major conclusions based on them in this paper. Figures 171–85, Plate V, are typical representatives of 10 genera of microspores. Plates I–IV are made up of examples of sixty species or varieties of spores, mostly megaspores, distributed through six genera.

DESCRIPTIVE TERMINOLOGY OF SPORES

Two basic requirements for any student or technician working with microfossil floras are the mastery of the understanding of the morphology of spores and pollen grains and the accumulation of an adequate, direct, simple vocabulary to discuss or describe each fossil. Even a basic terminology may at first appear to be too complicated, too detailed, or too picayune to be of practical value. However, a straightforward presentation of the terms, accompanied by simple diagrams, such as those given by J. M. Schopf,[8] should be readily understandable. Recently G. Erdtman, in chapter v of his comprehensive book, *An Introduction to Pollen Analysis*,[9] brought together a well-illustrated and clearly written description and glossary of pollen and spore morphology and terminology. A portion of the material given there is properly credited to the two other outstanding treatments of this subject by R. Potonie[10] and by R. P. Wodehouse.[11]

Spores usually occur in either of two principal forms. One form, represented by the spores shown in Figures 174 and 178 (Pl. V), is of a somewhat "beanlike," slightly elongate shape, i.e., bilaterally symmetrical. A slitlike opening, the monolete aperture, is evident on the upper surface. In contrast to these, radially symmetrical spores are distinctive (Pl. I, Fig. 1). They are usually marked on one surface with a triradiate scar or figure. This is called a "triradiate tetrad scar" because of its origin during the formation and the development of these spores in tetrahedral groups. Figures 72–75 show examples of tetrads of spores in their original tetrahedral groupings. When these groups are broken or

[8] "The Paleobotanical Significance of Plant Structure in Coal," *Trans. Ill. Acad. Sci.*, Vol. XXVII (1936), p. 109, Figs. *A–C*.

[9] Waltham, Mass.: Chronica Botanica Co., 1943.

[10] "Zur Mikrobotanik der Kohlen und ihrer Verwandten. I. Zur Morphologie der fossilen Pollen und Sporen," *Arb. Inst. f. Paläobot. u. Petrogr. d. Brennsteine*, Vol. IV (1934).

[11] *Pollen Grains* (New York: McGraw-Hill Book Co., 1935).

forced apart, the component spores appear individually as those single spores shown in Figures 66–70. Bilateral spores usually develop in linear or cruciform series of four, and such tetrads are rarely found intact.

Spores, either radial or bilateral, are generally oriented in a particular way for comparative study and description. The proximal (inner in relation to original grouping) and distal (outer) sides, or halves, of radially symmetrical spores are illustrated by Figures 66 and 69, respectively. The rays of the triradiate tetrad scar, which develop at the edges of the contact areas (portions of surface of a spore which were in contact with adjacent spores of the tetrad), may extend from the apex (center) to the equator (margin) of the proximal side (Figs. 87, 143) or partway to the equator (Figs. 14, 18, 185); or these rays may extend beyond the margin of the spore body onto a peripheral flange, a structure which is formed by an unusually great development of the arcuate ridge which sometimes borders the outer edge of each contact face (Fig. 182). The interradial contact areas (contact faces between rays of triradiate scar) may be called "pyramic areas." They often differ in surface texture from any remaining peripheral zone of the proximal half of the spore body proper. Figures 16, 18, and 19 show clearly the smooth pyramic areas of the proximal side with a well-developed triradiate scar, which is more prominent at the apex than toward the equator. Bordering the smooth pyramic area in these figures, the arcuate ridges are scarcely visible, but they are well shown at the right and left center of the spore in Figure 20 and on the spores illustrated by Figures 40, 58, 149, and 156. Peripheral to these ridges in Figures 16, 18, and 19 a knobby or verrucose ornamentation of the surface extends to the equator. This ornamentation continues over the distal surface also, as may be seen in Figure 17, which shows the opposite (distal) side of the same spores. In a few cases the arcuate ridges coincide with the equatorial rim and may be further developed into a flange, as has already been suggested. Such flanges may be smooth and thin or leathery (Fig. 143), striated in texture (Pl. I, Fig. 1), or delicately dissected (Pl. I, Figs. 2, 3). Sometimes the flange or arcuate ridges show particularly strong development at the ends of the rays of the triradiate scar or angles of the spore body. Figure 139 shows clearly this type of auriculate or "eared" spore. In the center of some radially symmetrical spores is a trilete aperture, an opening (Figs. 185, 14, 40) which corresponds in function to the aperture found on some bilateral spores as stated earlier. Such apertures, or the split developing along the trilete suture, are the points of emission of the plant developing from the spore (gametophyte). The apertures are often masked by extra development of the spore wall along the rays. In one group of spores (section Lagenicula) an exceptional development of such accessory lips is important in the formation of a prominent, somewhat pyramidal apex, called a vestibule (Figs. 10–12, 100–104). Such apical development results in bottle-shaped or lageniculate spores, in which the polar axis (proximodistal axis through apex) is longer than the equatorial axis.

Further orientation in terminology may be derived from a careful study of the skilfully executed drawings and diagrams illustrating the basic plan of each of the genera of spores recognized in the Schopf-Wilson-Bentall treatise.[12]

[12] J. M. Schopf, L. R. Wilson, and Ray Bentall, "An Annotated Synopsis of Paleozoic Fossil Spores

PRINCIPLES OF INTERREGIONAL CORRELATION

The megaspore flora of the Appalachian coals has been found to include a number of species recorded from the Upper Paleozoic deposits of Europe, as well as from the interior fields of North America. It is hazardous to suggest precise relationships of plants or isolated plant entities, found in geographically distant places. Modern floras, however, show a strong genetic relationship between plants of different regions or continents, and in many cases even species are known to be distributed widely over the earth. Sometimes their distributions appear to be more strongly controlled by habitat or life-zone environments than by continental limitations and such continental barriers as mountains and deserts. It is on the premise that the controlling factors for plant distribution are much the same today as at other periods of the earth's history that we venture to assume that fossil plants of the Coal Measures of Europe may have been specifically related to those of the same period in North America.

Caution, however, must be observed in assigning specific epithets to fossil plants which are apparently indistinguishable but are of different geologic distribution. Many genera of plants today have been traced directly back through Late and Middle Tertiary, and some even further. Only a few of the many thousands of existing species, however, are believed to have a continuous record, even from the Miocene to the present. With an appreciation, then, of the long period of time included in the Permo-Carboniferous period or in the Pennsylvanian alone, it is necessary to use great discretion in assigning to a spore from the Monongahela formation, for example, the same name given originally to a specimen from the Kanawha or Westphalian A.

SPECIFIC CHARACTERISTICS AND POSSIBLE VARIATIONS IN A GENUS

Plate I (Figs. 1–25) was assembled to show, comparatively, some of the variation possible in spores assigned to a single genus. This genus, *Triletes*, is the largest and most complex known at the present time. Schopf, Wilson, and Bentall[13] recognize 95 species to be included within it, some of which are of Mesozoic age.

Many of these are admittedly of doubtful status, some probably are conspecific with species already established, others may be of subspecific rank. Schopf, at the meeting of the Paleontological Society of America in 1936, proposed a division of *Triletes* into four groups or sections. Previously, other section names had been suggested, all but one of which (Lagenicula, Bennie and Kidston) appeared untenable to Schopf because of the unnatural relationships resultant from such grouping. He later restated these sectional divisions, Aphanozonati, Lagenicula, Auriculati, and Triangulati, and appended a description to each.[14]

At the top of the chart in Figure 2 the list of 60 kinds of spores includes 54 presumed to belong to *Triletes*. They have been grouped there, as nearly as possible, into the sections described by Schopf. In each group the species and varieties have been given in alphabetical order, and each name is referred by number to its illustration(s) in Plates I–IV.

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and the Definition of Generic Groups," *Ill. Geol. Surv., Rept. Investigations No. 91* (1944), Pls. I–III.

[13] *Ibid.*, pp. 18–27.

[14] Pp. 17–24, 27, 30, 31 of ftn. 7.

The range in size is suggested by a comparison of (Pl. I, Fig. 1) one of the largest megaspores with Figure 6. These species, *T. superbus* and *T. triangulatus*, respectively, both have prominent equatorial appendages. The spore body of the former, therefore, is not so large as *T. reinschi* (Figs. 28–40). The largest megaspores belong to the genus *Cystosporites*, which is not figured here. They range up to 2–3 cm. in length. All others are generally less than 3 mm. in diameter. The smallest that can be conveniently picked out of the residues by manual manipulation are of about 300μ in diameter.

Equatorial or zonal appendages are prominent features of such spores as those shown in Plate I, Figures 1–3 and 6, but are less prominent in auriculate forms, such as those shown in Figures 4, 9, and 13–15. Often the flange of a spore may be broken off before or during fossilization or in maceration. Therefore, the intimate characteristics of the spore body itself must be recognized for proper identification. Figure 22 shows the spore body of *T. brasserti*, with the flange missing. There may be considerable variation in width of flanges in the same species, as shown by a comparison of *T. rotatus* in Plate I, Figures 2 and 3, and in degree of dissection. It is extremely difficult, if not impossible, to draw the line between species which have been founded solely on variable characteristics of this sort.

Variations of a triradiate tetrad scar are illustrated by the series of Figures 8, 16, 19, 3, 1, and 22 (Pl. I), in that order of increasing prominence, and Figures 14, 5, 8, 16, and 3, in the order of increasing length of rays. Apical structures of increasing prominence are shown in the comparison of *T. tenuispinosus* in Figures 16, 18, and 19 with *T. levis* in Figure 12 and *T. kidstoni* in Figures 10 and 11. The strongest development of this type is shown by such a species as *T. nudus* in Figures 100–103.

All the foregoing features are presumed to be genetically derived in the development of the spore and may be termed "haptotypic" characters. Genetically determined characters should be reasonably constant and should afford bases for separation of organisms into major groups.

Such features as surface ornamentation, however, appear inconstant and can be used only for varietal or perhaps specific distinctions. Schopf[15] has illustrated a practical terminology for such emphytic characters. Figures 12 and 23 show a smooth or levigate surface. The pyramic areas of *T. tenuispinosus* in Figures 16 and 18 are finely punctate or granulose; in *T. brasserti* (Fig. 22) and *T. superbus* (Pl. I, Fig. 1) they appear verrucose and verrucose to spinose. Toward the edge of the pyramic area in Plate I, Figure 1, the surface is covered with hairlike protrusions which peripherally grade insensibly into the rami of the flange. Actually, the flange *appears* to be made up of a dense mat of these hairs, irregularly fused together. The pyramic area in *T. praetextus* (Fig. 20) is very rough or verrucose, much the same as the rest of the spore surface, but in Figures 16–19 the granulose texture of the pyramic area is markedly differentiated from the spinose spore body. In *T. kidstoni* (Fig. 11) the spines are hooked, more slender, and less numerous, whereas in *T. tuberosus* (Figs. 24 and 25) they are strong and tuberculate.

Sporelike bodies similar to those shown in Figures 26 and 27 have been found in widely separated horizons of the Pennsylvanian. They vary greatly in

[15] Explanation of Fig. *C*, ftn. 8.

size, but, by the nature of the reticulation on the inner surface of the transparent wall, they appear to be sporangia, i.e., the spore-bearing sacs, possibly of certain calamitean plants.

Differences in position and nature of preservation, in maceration, and in conditions of observation may result in apparent variations of spore bodies. Figures 4, 5, 7–9, and 13–15 may be compared in these respects. Figures 7, 9, and 14 show *T. auritus* compressed axially. Examples of this species in Figures 13 and 15 are compressed in an equatorial plane, thus showing parts of both proximal and distal halves. Figure 9 is brightly illuminated, and reflection from the dense black surface makes the spore appear light and nearly obscures the prominent infolded "ear" on the upper right-hand margin. Figure 13 is illuminated satisfactorily and shows the arcuate ridge (at left) between the ends of two rays of the trilete mark which converge at the apex (lower right). Figure 15 is very weakly illuminated but shows the dark color and granular texture of the surface. The arcuate ridges extend from upper left to lower right; and, since in this species the arcuate ridges nearly coincide with the equatorial rim, it may be said that the lower left portion of the picture is part of the proximal half and that the upper right (crescent-shaped) portion is distal. One ray of the trilete mark extends across the proximal area and is concealed at its peripheral (equatorial) extremity by a boss or knob which is actually an "ear" of this auriculate form. The "ears" may be preserved in an extended condition (lower right of Fig. 5) or folded over the body as already mentioned (upper right of Fig. 5). The pyramic area of Figure 7 shows some roughened areas, which are patches of weakly developed spines; but in Figures 4 and 5, on a variety of this species, *T. auritus grandis*, the roughenings are caused by extraneous matter adhering to the spore, an indication of undermaceration. The trilete aperture in Figure 14 is open, and the spore shown in Figure 8 would present the same type of opening if it were split along the rays of the trilete figure. Such variation in openings appears to be due to various happenings either before or during preservation, compression, or maceration.

Such instances of variation among different specimens of the same species could be multiplied indefinitely. This variability is the most difficult problem for the student of such fossils. It is the basic reason for the confusion which exists in their taxonomy, for the actual *specific value* of the variations has not been determined.

DISCUSSION OF THE GENUS *Triletes*, SECTION APHANOZONATI

All the figures on Plate II represent species of the section Aphanozonati, with the exception of two spores of uncertain relationships (Figs. 59–60), referred to *Sporites plicatus*. They are believed to be spores of certain large lycopods, because many similar spores have been isolated from spore-bearing structures (strobili) of *Lepidodendron* and *Sigillaria*.

This group is typified by *T. reinschi* (Figs. 28–40). Aphanozonate spores lack flanges and other equatorial appendages, but arcuate ridges are normally evident (Figs. 40, 56, 58). The spore coat is usually thick and has considerable variation in ornamentation, from smooth (Figs. 33, 32) to spiny (Figs. 46, 51) or knobby (Fig. 47). Included in this group are some of the largest megaspores, ranging in diameter up to about 3 mm. The triradiate mark (Fig. 33) or aperture (Figs. 29, 37, 40, 42, and 46) is usually restricted to

less than one-half the diameter of the proximal surface.

T. reinschi (IBRAHIM) SCHOPF (FIGS. 28–40)

This species is fairly common throughout the Pennsylvanian coals of the United States. It is regarded by Schopf as having a long range in the coals of Illinois; and in the Appalachian region it is found from the Gilbert coal (near the base of the Kanawha group) to the Waynesburg. It is locally abundant in several seams.

It is a large spore, varying from 1,700 to 2,700 μ, usually circular in outline and flattened axially. The surface is rather smooth and varies from dull black to light brown. The pyramic area is defined by a low arcuate ridge, which is often difficult to discern. The rays of the aperture are slender, extending one-third to one-half the radius of the spore, and they end in an acuminate tip. They are often split open along these sutures (Figs. 29, 37). The interradial areas of the spore are occasionally plicate. These folds are irregular but extend radially from the apex.

In some instances the distal surface has a few short tubercles irregularly distributed. These are rarely found on the proximal surface and never in the pyramic area. Other forms appear minutely pustulose. These types may actually be varietally or specifially different.

Two varieties of special nature are shown. Figure 43 is of a form found in the Waynesburg coal. It seems to have a more conspicuous pyramic area, which appears convex in flattened state. The first of these found was presumed to have had the outline of another spore impressed on it, coinciding with the pyramic area, but subsequently discovered specimens make such a speculation improbable.

Figures 53 and 61 are of spores from the Eagle coal. They are usually folded, somewhat triangular in outline and smaller than, but otherwise similar to, *T. reinschi*. There are no hairs or peculiar markings, but the surface is finely granulose. It does not appear to be assignable to *T. fulgens* or *T. glabratus* but may be a small *T. reinschi*.

T. apiculatus (IBRAHIM) SCHOPF, WILSON, AND BENTALL (FIGS. 46, 51–52)

A relatively large, gibbous spore, the *T. apiculatus* is irregular but not angular in outline and is 1,500 μ in average diameter. The body is covered both distally and proximally, except in the restricted pyramic area, by large papillae or apiculae. These are higher than broad, distally bent or recurved, and terminated by sharp points. The pyramic area is demarcated principally by lack of these large apiculae, though numerous short spines or pustules are present. The arcuate ridge at the periphery of this area is very inconspicuous, but it does completely outline the area. The rays of the trilete aperture are small and narrow, usually extending not more than one-third the distance to the periphery of the proximal side. They form no prominence at their juncture in the center. The exine appears to be thick and opaque.

It has similarities to *T. brevispiculus* Schopf and to *T. mamillarius* Bartlett and to some specimens of the Type 14 of Zerndt. This species is recorded from Namurian and younger rocks of Europe and has not been found above the Kanawha group here.

Three varieties are also illustrated. One variety, *T. apiculatus* var. I (Figs. 41 and 42), is a giant form, 3 mm. in diameter. It is sparsely spinose, and very long, fine hairs are conspicuous along the rays. Portions of the surface are pustu-

lose. It appears to be intermediate between the unornamented *T. reinschi* type and *T. apiculatus sensu stricto*. The second, *T. apiculatus* var. II (Fig. 44), is another form intermediate between *T. reinschi* and *T. apiculatus*. In this variety the pyramic area is greatly reduced, and the trilete scar is distinct but delicate. The interradial areas are finely folded (plicate). Variety III (Figs. 48–49) is only 1,100 μ in average diameter, with long trilete rays, which are developed prominently at the apex. The coat is rough but not knobby, and true hairs are lacking.

T. brevispiculus SCHOPF (FIG. 45)

This is a moderately thin-walled, light-brown megaspore, with scattered mammillate papillae. This species, so far found only in the Upper Freeport coal, coincides remarkably with the description given by Schopf.[16] The pyramic area does not show in this photograph, for the nature and distribution of spines over the tan surface were deemed most important for illustration. An unusual instance of correlation is afforded by this species, for Schopf's material is from the No. 6 (Herrin) coal of Illinois, near the top of the Carbondale series. The position of this coal, as may be seen on the National Research Council's correlation chart,[17] is presumed to be close to that of the Upper Freeport, which is near the top of the Allegheny formation. Additional studies may extend the range of this species in either or both coal fields.

T. mamillarius BARTLETT (FIG. 47)

The spore is roughly globose, prominently marked by very strong, mammilliform papillae. The specimen shown here, which is from the Winifrede coal (Upper Kanawha), is broken. The spine bases show most clearly near the edge of the pyramic area, which is not well defined. A number of other species from Europe are closely related. The stratigraphic position of this and some other spores from Upper Pottsville coals of West Virginia may assist not only in placing the drift pebbles of coal, described by Bartlett simultaneously with the description of this and two other megaspores,[18] but also in establishing correlation of the Michigan coal seams with the Appalachian coals.

T. fulgens (?) ZERNDT (FIGS. 50, 56–57)

The identification of the spore types shown in Figures 50, 56, and 57 is not final. They have nearly twice the average diameter given for the species. Otherwise, Zerndt's figures[19] are certainly similar in general appearance. In the best specimen (Fig. 56) from the Pittsburgh coal horizon, the surface is dull and slightly roughened proximally and satin-smooth and black, distally. The form is somewhat gibbous, and strong arcuate ridges surround the pyramic area. There are prominent interradial plications.

The considerably younger age (Monongahela) of this material than Zerndt's, which is Middle Dinantian to Middle Namurian, in addition to difference in size, may be sufficient for specific differentiation.

[16] Pp. 26–27 and Pl. I, Figs. 13*a–r;* Pl. II, Fig. 6; Pl. III, Figs. 1–4, of ftn. 7.

[17] Pennsylvanian Subcommittee, "Chart No. 6. Correlation of the Pennsylvanian Formations of North America," *Bull. Geol. Soc. Amer.*, Vol. LV (1944).

[18] Harley Harris Bartlett, "Fossils of the Carboniferous Coal Pebbles of the Glacial Drift at Ann Arbor," *Papers Mich. Acad. Sci. Arts, Letters, 1928*, Vol. IX (1929), pp. 11–25, Pls. XIII–XV.

[19] Jan Zerndt, "Les Megaspores du bassin houiller polonais. II," *Acad. Polon. Sci. Lett. Trav. Geol. No. 3* (1937), p. 5, Pl. I, Fig. 2; Figs. 1–9; and "Megasporen als Leitfossilien des produktiven Karbons," *Acad. Polon. Sci. Lett. Bull. Internat.*, Ser. B (1931), p. 171, Pl. III, Figs. 9–10.

One variation shown here, *T. fulgens* var. II (Figs. 54 and 55), is a little smaller and lighter in color, with a thinner spore coat and much less conspicuous arcuate ridges. In some respects this more nearly resembles Zerndt's material than those described above. Figure 55 shows a specimen flattened equatorially, with the spore broken open (at top of picture) along the trilete process.

A second variation, to be found in Plate IV, Figures 160–61, is about 600 μ in diameter with features very similar to variety II, above, except that the pyramic area occupies nearly all the proximal surface.

T. glabratus ZERNDT (FIG. 58)

This is a rather distinctive spore, which is of the same general plan as *T. fulgens*, but it has a number of characteristic features of specific value. This figure is closely comparable to Figures 2 and 3 of spore Type 9 in Zerndt's "Megasporen aus einem Flöz in Libiaz (Stéphanien)."[20]

Triletes glabratus is a very black, gibbous spore, with some roughening apparent on the dull surface. In the equatorial region it is slightly and finely pustulose, with slender, short hairs. The arcuate ridges are accentuated by the thinness of the wall in the pyramic area, which thus appears sunken. The trilete rays converge at the apex to form a small boss, and they expand radially, gradually blending on either side with the arcuate ridges. They often falsely appear to be extended beyond these ridges because of the presence of a broad fold or sinus, terminal to each ray. This species, from the Upper Kanawha, occurs approximately at the level of the base of the long European range given by Zerndt.

[20] *Ibid.* (1930), Pl. I.

T. simplex (ZERNDT) SCHOPF, WILSON, AND BENTALL (FIG. 134)

The large (1,700 μ) coarse spore has a rough, dark-brown coat, which is somewhat convoluted. The tumorous structure toward the left in the figure is the robust apex. The arcuate ridges are relatively weak. It is a peculiar fossil, with information lacking for systematic assignment. It has been found only in the generally poor residues from the oldest coals of the Kanawha.

Triletes simplex var. I is a larger spore with a thinner wall. Plate IV, Figure 133, shows two of a group of eight or ten which were pressed together in a mass isolated from the coal residue. The surface, in marked contrast to the rough, lusterless coat of *T. simplex*, has a vitreous luster and a minutely pitted surface. The walls here are moderately thin and much folded, and the trilete apparatus is usually obscured by the folds.

Triletes simplex var. II is a problematical variety (Figs. 131–32) and is shown here to be somewhat folded and with a thicker wall than variety I. The coat is tan and moderately translucent. These two specimens were recovered from the Pittsburgh and Winifrede coals, respectively.

T. kidstoni (LOOSE) SCHOPF, WILSON, AND BENTALL (FIGS. 10–11, 119–20)

The strong apical development apparent in Figures 10 and 11 is not typical for the species. The two specimens in Figures 119 and 120 show at upper left and upper right, respectively, a light, clear zone, which is the pyramic area. The spores are densely to moderately covered with thin, hooked spines. These two spores are more representative of the species, and they also show its comparatively smaller size than most of the

Aphanozonati. This species is very abundant in Upper Pottsville coals and often outnumbers all others in some benches. There are apparently several varieties, one of which may be represented by Figures 10 and 11.

Two forms, here considered to be at least of varietal status (*T. kidstoni* var. I and var. II), are from the Lower War Eagle coal. No intermediate occurrence of this type of spore between this horizon and Upper Pottsville is known. Variety II has strong arcuate ridges, with a slight thickening at the angles of the spore. Mammilliform papillae are loosely scattered over the surface, except for the extremely smooth pyramic area. As in *T. kidstoni*, this area is restricted to about one-third to one-half the proximal surface. Variety I is larger and has a stronger development of the apical processes. In fact, it is quite possible that a spore with the features shown in Figure 125 belongs to Lagenicula. These spores are among the best coal-index fossils of all the megaspores studied so far, because of their local abundance and apparent restrictions of geologic range. Additional work is required to determine their specific limitations.

T. tuberosus (IBRAHIM) SCHOPF, WILSON, AND BENTALL (FIGS. 24–25)

The species *T. tuberosus* was partially described earlier. It is another of the very strongly apiculate forms. The demarcation between this and several others is indefinite. It is exceedingly abundant immediately below the Winifrede limestone horizon (Upper Kanawha) in all the beds of the multiple Chilton coal seams.

T. tenuispinosus ZERNDT (FIGS. 16–19)

The features and characteristics of *T. tenuispinosus* have been discussed and illustrated in connection with the study of Plate I.

Sporites plicatus SCHOPF (FIGS. 59–60)

Although this problematical form is discussed at this point, it is not to be inferred that it is considered to belong to the Aphanozonati. It could be placed at the end of the entire résumé of spores but is treated here because of its inclusion in Plate II.

There are a number of spores which show such generalized features that they cannot be assigned to any group. A spore assigned to the genus *Sporites*, therefore, has little biological significance and will probably never be an aid to correlation. If the spores included in this "form genus" eventually become better understood, new groups will probably be made to receive them. Species assigned to this genus are not presumed to be related in any way. Apical structures of these spores, as shown in the figures, are usually obscured. They may have either a trilete or a monolete suture or neither. The wall of this "species" is folded and slightly translucent. Representatives of this heterogeneous group occur throughout the Pennsylvanian coals.

DISCUSSION OF THE GENUS *Triletes*, SECTION TRIANGULATI

The basic characterization of this group is less distinctive than desirable, and the determination of the spores to be included is difficult. Plate III was assembled to show some of the more conspicuous types. Figures 62–79 and 84–106 include spores of this group with the most conspicuous zonal appendages. In Plate IV several additional species are illustrated in Figures 111–14, 135, 140–41, and 155.

The name "Triangulati" is derived

from the somewhat triangular appearance of certain of these spores, an outline due to the unequal extension of the flange about an oval or circular spore body in some species. A number of spores included in this group are not so characterized. The pyramic area is usually large, and the arcuate ridges and other appendages are equatorial (or nearly so) in position. The trilete rays are usually quite long.

Schopf[21] suggests that spores assigned to the Triangulati represent such herbaceous lycopods as Lycopodites and Selaginellites. He believes that these groups and allied types are probably more important in coal formation than is usually recognized from macroscopic remains. They certainly have a prominent position in coals of the Upper Pottsville in the Appalachian field, although they are much less important, i.e., have fewer representatives, at earlier and later periods, according to an analysis of the spores studied to date. Two or three species(?) have extremely long ranges, however.

T. superbus BARTLETT (PL. I, FIG. 1; PL. III, FIGS. 62–65)

Several spores are so closely allied to this species that specific determination is difficult. Surely one or more other species may be conspecific with it. A detailed discussion of this species will suffice at this time for most of these closely allied spores.

Bartlett concluded his original description[22] of *T. superbus* with the statement that it was so named ". . . . because it is unsurpassed in size and beauty of form among all the spores that have been described from coal." His early paper well illustrates the striking features which characterize *T. superbus*, though the size is not so great as that of some other Paleozoic spores, as previously noted. Figure 1 (Pl. I) has been described in part. The body of the subtriangular spore is about 2,200 μ in diameter. The rays of the strong triradiate structure are cleft longitudinally and extend to the periphery. Each ray appears as a double band of "leathery fringe" standing on edge on the spore surface. The flange is composed of a convoluted series of fused cylindriform processes (rami), more thickly matted near the inner margin than peripherally. The entire spore body is usually covered with these hairlike processes, though toward the apex they are reduced in size and scattered less densely.

T. brasserti STACH AND ZERNDT (FIGS. 22, 66–77)

Triletes brasserti is a spore similar to *T. superbus*, with slightly coarser features. Figure 22 on Plate I shows clearly the exceedingly prominent bi-ridged rays of the trilete process, which is more strongly developed but otherwise similar to the preceding species. The distal side is essentially smooth (Fig. 69), but, toward the periphery, minute protuberances occur, which become larger peripherally to blend with rami of the flange. The flange is usually less than one-half the radius in width. For orientation, it may be noted that on the tetrads shown in Figures 72–75 only the distal side of each component spore is visible and that the flanges are developed at the margins of contact with the concealed faces of each adjacent spore. Figures 68, 76, and 77 show spores which were compressed equatorially instead of in the normal axial plane.

The spores portrayed in Figures 87–90 are probably a distinct variety. They are

[21] P. 32 of ftn. 7.

[22] Pp. 20–21 of ftn. 18.

usually smaller, with a very narrow flange. Another distinction is the slightly reduced pyramic area, a condition well shown in Figure 90. The flange is not quite coincidental in position with the equator.

T. circumtextus ZERNDT (FIGS. 78–79)

Triletes circumtextus is closely related to *T. brasserti* but is a smaller spore with more delicate features. It is closely comparable to the Type 18 of Zerndt, especially 18*a*, 18*c*, and 18*d* of his 1934 paper on "Les Megaspores du bassin houiller polonais."[23] The flange varies from one-third to one-sixth the diameter of the spore and from solid to reticulate or lacerate. It appears to be composed of anastomosed radial plates or fibers.

The surface is dull and minutely papillose in the pyramic area, and on the distal surface the papillae are larger, more numerous toward the periphery, and absent toward the center. This species has a slightly later range than *T. brasserti*, i.e., it is prominently developed in the Lower Allegheny, and appears to be the only one of these closely related species to extend so high in the section, according to present records of observation.

T. rotatus BARTLETT (PL. I, FIGS. 2–3, 84–86)

The most conspicuous feature of this spore is the unusually delicate structure of the flange. It is composed of a number of slender, sinuous processes which radiate from the spore body at wide intervals and fuse together by lateral branches at the extreme periphery to form a narrow rim. Figure 3 is scarcely typical because of the more numerous radiating processes and the greater degree of fusion. Figure 2 (Pl. I) is atypical because of the unusual narrowness of the delicate flange. Figures 84 and 85 are poor pictures because too much was attempted by illuminating both background and spore surface. Normally, the flange is torn off to a greater or lesser extent, though, with careful handling and undermaceration, some complete specimens can be obtained.

Transition is complete from a type with extremely delicate flange with unfused, ctenate (comblike) processes to ones with elements terminally fused to form a rim, and to forms with a reticulate flange. The actual specific value of the flange type is unknown. This alliance includes species or varieties which are among the earliest of the triangulates, appearing even in the Eagle coal. One species, *T. radiatus* (Ibrahim) Schopf, Wilson, and Bentall, which will not be considered here separately, is of considerably later range, principally Allegheny, in the Appalachian coals. It may be that Figure 3 is of this type. The author has examined large quantities of some megaspores from the Michigan coal basin, including the *T. rotatus* material, but at present is unwilling to differentiate that species *sensu stricto* from gradational forms including *T. radiatus*.

T. praetextus ZERNDT (FIG. 20)

The coarse-featured spore in Figure 20 was described in the discussion of Plate I. The arcuate ridge is prominent but can scarcely be considered a flange. The diffuse pustules or stubby spines vary considerably in different specimens. The most common occurrence of this species is in the mid-Kanawha. The age of the European material is Upper Dinantian and Lower Namurian. Such a stratigraphic discrepancy may be one evidence of specific difference of the Appalachian material.

[23] Pp. 19–21, Pls. XIX, XXII, and XXIII of Part I (1934), of ftn. 19 (1937).

A variety shown in Figures 91–94 is considerably smaller; and the thinner, wavy, fringelike arcuate ridge may be interpreted as a very narrow, solid flange. The rays of the triradiate tetrad scar are prominent, ribbon-like, and irregularly fluted. This form somewhat resembles *T. hirsutus* but does not appear to be lageniculate. It is important in the Lower Kanawha coals.

T. clavatopilosus (WICHER) SCHOPF, WILSON, AND BENTALL (FIGS. 95–99)

The examination of spores of *T. clavatopilosus* is rendered difficult by the amount of extraneous residue which appears to be matted over the surface, together with the filiform appendages. It may be that the specimens shown are incomplete. The hairs seem to be short and appear indistinguishable from the rami of the lacerate flange. Spores such as this from the lower coals are difficult to recover in good condition. A study of them is aided by an examination of thin sections.

The spore shown in Figure 135, Plate IV, is possibly related. The trilete structure is nearly the same, i.e., very strong and leathery in appearance, with considerable apical development. The flange, however, is narrower and made up of more delicate protuberances.

Three additional species presumed to belong to the section Triangulati which are not shown on Plate III are discussed here to avoid discontinuity.

T. triangulatus ZERNDT (FIGS. 5 AND 112–14)

The general appearance of the spore is best shown by Figure 5, already discussed. It is one of the most common of all megaspores and has a range throughout the Middle and Upper Pennsylvanian. It may be considered typical for the section Triangulati. Schopf[24] has discussed this species in considerable detail. One spore feature, not mentioned in the preceding pages, is the presence of a network of thickenings (reticulation), which occurs on the wall of the spore body and is notable in this species. The reticulation of the distal surface is more prominent than that of the proximal half. Occasionally, a system of reticulation is to be found on the flange. There is a considerable range of size (300–800 μ) and ornamentation on specimens assigned to *T. triangulatus* which may be of specific rather than varietal quality. In the variety shown in Figure 111, the flange is thicker and perhaps wider, and the trilete rays are shorter.

T. artecollatus NOWAK AND ZERNDT (FIGS. 140–41)

Triletes artecollatus appears to be somewhat ambiguous. It may be more reasonably associated with the Auriculati. The spores shown here more closely resemble the figure given by Zerndt[25] for his Type 39 than they do the original description and figures.[26] They have a pustulose surface, weak trilete apparatus, and strong arcuate ridges. The surface of the pyramic area is almost "warty" near the apex. The much younger age (Allegheny) of this material than the original specimens (Dinantian) makes specific identification questionable.

T. gymnozonatus SCHOPF (FIG. 155)

The final species of the Triangulati discussed here is a small, unornamented, smooth, black megaspore. The arcuate

[24] Pp. 32–37 of ftn. 7.

[25] Fig. 11, p. 16, of ftn. 19 (1937).

[26] J. Nowak and Jan Zerndt, "Zur Tektonik des östlichsten Teils des polnischen Steinkohlenbeckens," *Acad. Polon. Sci. Lett. Bull. Internat.*, Ser. A (1936), p. 58, Pl. I, Figs. 4–5.

ridges are present but inconspicuous. It is identified with reasonable certainty, and the age of the specimens found is Upper Allegheny (Upper Freeport coal). Here again is a fortunate specific correlation between spores of approximately the same age in the Appalachian and eastern Interior coal basins. Megaspore floras of more coals from the Illinois Basin will have to be studied to determine the relative value of such spore correlations.

DISCUSSION OF THE GENUS *Triletes*, SECTION AURICULATI

The spores of the Auriculati are characterized by the radial development of lobes (auriculae) opposite the trilete rays. The flange is usually weakly developed at the periphery of the interradial areas.

T. appendiculatus MASLANKIEWICZOWA

The three-lobed character of spores of the group is inconspicuous in some species, but in this one, from the uppermost beds of the Kanawha or Lower Allegheny, the lobes are particularly prominent. An illustration of this interesting form was inadvertently omitted. It has auriculae which extend a distance of nearly one-half the radius of the spore beyond the spore body, and on these ear-like flaps are short, sharp spines or tubercles. The spore body is granulose, and the trilete apparatus is very thick, with the rays tapering to sharp points. A gradational series from species of this sort to species with inconspicuous "ears" occurs.

T. auritus ZERNDT (FIGS. 7–9, 13–15, 151–52)

This species has been described in the earlier discussion of possible specific variations. It may be differentiated from its varietal forms only with difficulty. The variety *T. auritus grandis* Zerndt is presumed to be represented by the form with bulbose "ears" shown in Figures 4 and 5. Variety II (Figs. 147–50) may be favorably compared with *T. auritus* in the adjacent figures (Figs. 151–52). Such haptotypic features as flange and trilete apparatus are recognizably different in the two. The spores in this group are somewhat similar to some of those described as *T. silvanus*. The specimen of *T. silvanus* illustrated in Ibrahim's dissertation[27] probably is less conspicuously flanged, and the auriculae are smaller, but in the material illustrated by Wicher[28] as *Sporites silvanus* (Ibrahim), the flange and the auriculae are closely comparable to those of *T. auritus*.

Variety III (Fig. 156) and variety IV (Figs. 157–59) are poorly illustrated. They are about one-third the size of *T. auritus* and are smooth and densely black. The auriculae in variety IV are larger, and the flange or arcuate ridge is less conspicuous.

T. silvanus (IBRAHIM) SCHOPF, WILSON, AND BENTALL (FIGS. 142–46)

The species *T. silvanus* is somewhat similar to *T. auritus*, just described. The specimens illustrated are larger than the type material and perhaps should be specifically differentiated. The coat is usually granulose, though the proximal side is roughened in some specimens, and on the distal side it may be warty. The flange is a convoluted, leathery fringe coinciding with the equator. The pyramic area is divided by a strong trilete apparatus with high, wide, crenulated

[27] Ahmet Can Ibrahim, *Sporenformen des Aegirhorizonts des Ruhr-Reviers* (Private printing; Wurzburg: Konrad Triltsch, 1933), Pl. II, Fig. 22.

[28] Carl A. Wicher, "Sporenformen der Flammkohle des Ruhrgebietes," *Arb. Inst. f. Paläobot. u. Petrogr. d. Brennsteine*, Vol. IV (1934), Pl. VIII, Figs. 5–8.

lips. In general character the spore is coarse and usually about 2,000 μ in diameter. It occurs in the Monongahela series, relatively higher than the European specimens.

T. nigrozonales STACH AND ZERNDT (FIGS. 136–39)

The species *T. nigrozonales* is quite dark and smooth, with sharply defined haptotypic features. The flange is narrow and slightly crenulated. The auriculae are strongly developed. It appears to be distinct from a number of other Auriculati which occur with it in Lower Allegheny and Upper Kanawha beds. Its range is short and comes near the base of the short European range (Westphalian C).

DISCUSSION OF THE GENUS *Triletes*, SECTION LAGENICULA

The chief characteristic of the lageniculate group of *Triletes* is the remarkable development of the apical segments of the spore coat to form a vestibule. The resultant shape of the spores has been discussed. There are a number of forms which appear to be intermediate between these and some of the other spores already described. Schopf[29] discussed the difficulty of determining the affinity of spores to Lagenicula by the presence of the vestibule alone.

T. crassiaculeatus (ZERNDT) SCHOPF, WILSON, AND BENTALL (FIGS. 107–10)

The spores shown here may differ from the type material in several features, but no other specific characterization fits these as well. The vestibule is slender and small but well differentiated. The surface is sparsely spinose. The spore coat and surface resemble those features of *T. kidstoni*. It has been recognized only from some of the older coals of the Kanawha group.

T. hirsutus (?) (LOOSE) SCHOPF, WILSON, AND BENTALL (FIGS. 80–83)

This species is not positively identified. The material illustrated here is older than European specimens. The spore body, about 800 μ in diameter, somewhat resembles that of rotate spores, such as *T. radiatus*. The distal side is covered by an indument of cylindriform rami of varying lengths. Individual rami extending radially from the equator may be the sole remains of the lacerate type of broad flange.

T. levis (ZERNDT) SCHOPF, WILSON, AND BENTALL (FIG. 127)

Several spore types have been found which are tentatively considered to belong to this species and its varieties. The specific distinction of relatively unornamented Lageniculae is difficult. *Triletes levis* of the European fields is Stephanian in age. Varieties I and III (Figs. 126 and 130, respectively) have been found in the Waynesburg coal and some of the Lower Permian coals above, whereas the form here called *T. levis* and also *T. levis* var. II are found only in early Kanawha coals.

These spores have dense-black coats; thick, fleshy lips; and no evidence of arcuate ridges, flanges, or spines. The trilete apparatus is prominent, and the interradial plications are distinctive. The vestibule is prominent to inconspicuous in the different varieties. In Figure 127, *T. levis* actually looks very nearly like the much larger *T. simplex* in Figure 134.

Statistical analyses of additional material will probably result in a separation of some of the species based on early Kanawha forms from those of Upper Monongahela age.

[29] Pp. 27–28 of ftn. 7.

T. nudus (NOWAK AND ZERNDT) SCHOPF (FIGS. 100–104)

The five specimens shown in Figures 100–104 are all from the Waynesburg coal. This species has not been found in the younger seams of the Pottsville. The range of the type material is equivalent to Upper Mississippian and Pocahontas, as nearly as can be determined. Even with this discrepancy in age, the spores are indistinguishable on the basis of published descriptions and figures. If they are actually specifically different from the European material, criteria for so adjudging them are not recognized. Schopf has described some specimens from No. 6 coal of Illinois which may also be specifically distinct. The spores illustrated here show a well-defined arcuate ridge and very high vestibule, but no other surface appendages or emphytic ornamentation.

Very similar is *T. nudus* var. I, except that the arcuate ridge is inconspicuous and the spore in over-all appearance is gourd-shaped rather than bottle-shaped.

T. rugosus (LOOSE) SCHOPF (FIG. 105)

The figure is too small to reveal much of the nature of the spore. It has low, distinct lips; thick coat; rough, warty surface, with well-defined arcuate ridges and trilete apparatus. No vestibule is evident. Here again is a spore in the Upper Freeport coal which is approximately of the same age as the material found in the Illinois Basin.

T. translucens SCHOPF (FIGS. 115–16, 124)

Triletes translucens is somewhat similar in general character to *T. rugosus*. It has thinner walls and is usually smaller, although a disproportionate range in size is shown by comparing Figure 115 with Figure 116. These spores were carefully examined, and no significant differences other than size are apparent. The type shown in Figures 117–18 is recorded as a variety, but actually so little is determinable from the specimens found that it is of little value taxonomically and stratigraphically as it now stands. These spores can best be studied by transmitted light, whereas most of the preceding spores are examined principally by direct light.

Since no illustrations are given, no discussion of the other genera of megaspores will be undertaken at this time. No other genus has as many species as *Triletes*, but several species of each of several other genera have been found and are used for stratigraphic correlations.

DISCUSSION OF THE MICROSPORE ASSEMBLAGES

The microspores of the Appalachian coals have not been studied systematically. A total of fifty-eight species or varieties, however, are recorded from about twenty coal seams. Plate V shows fifteen different species in ten genera which are commonly found. All occur in three coal seams of southern West Virginia. These three seams—the Powellton, Campbell Creek, and Alma, all of mid-Kanawha age—are easily differentiated from one another in that area, on the basis of microspores. The Powellton shows an abundance of *Laevigato-sporites* (Figs. 174, 178) and *Lycospora* (Fig. 184). Very few other microspores occur. The Campbell Creek coal has a large number of microspores, including *Laevigato-sporites* and *Denso-sporites* (Figs. 177, 179), but it is particularly marked by several species of *Granulati-sporites* (Figs. 172, 173, 175). The Alma coal microspore flora is dominated by a certain species of *Denso-sporites*. Geographic variation of these floras is undetermined at this time. The identification of the material is impeded,

as it is in the case of the megaspores, by the confused status of the recorded species. Of the fifty-eight forms found, more than two-thirds appear to be new. From a preliminary analysis it appears that the microspores are more suitable than megaspores for distinguishing closely spaced coal seams and in identifying benches which may diverge from one another as splits.

MICROFOSSIL EVIDENCE FOR THE DIVISION OF THE PENNSYLVANIAN STRATA OF THE APPALACHIAN REGION

Examination of the microfossil flora in several geographically spaced samples from a single seam of coal shows, in many cases, a reasonable lateral variation of the species present. In most seams, however, a typical microfossil assemblage is identifiable for a major portion of the range. A detailed study is now being made of the floras present in five coal seams (Waynesburg, Uniontown, Pittsburgh, Chilton, and Campbell Creek). This should ultimately determine the reliability of typical spore assemblages and the amount of lateral variation which may be expected.

Evolutionary floral changes, which mark "breaks" in the geologic succession, are quite noticeable in studying the microfossils of successive beds. A conspicuous change appears in the flora above the Lower War Eagle coal, probably having occurred during the time of deposition of the Eagle shales and limestone. The floral assemblages of the subsequent coals (Cedar, Eagle, and others), up to the shales (including the Stockton limestone) above the Matewan coal, are rather uniform in the samples studied. The Powellton, Campbell Creek, Alma, and Cedar Grove coals, deposited in that order above the Matewan (?) shales, are remarkably uniform in megaspore assemblages. The prominently flanged spores of the Triangulati are most abundant and become progressively varied and specialized from the Powellton to the Cedar Grove. A marked change in the floras occurs somewhere in the interval before the deposition of the Hernshaw coal. The actual "break" may be explained by change of conditions represented by the Dingess shale and limestone series.

One of the biggest changes in the floras occurs between the Buffalo Creek coal and the later Coalburg. Even the admittedly sketchy distribution record given on the chart in Figure 2 shows a marked difference between the megaspores of these two coals. The principal relationship of the Coalburg megaspores to previous floras is found in a few vestiges of the highly specialized, flanged megaspores of the Triangulati. According to present knowledge, these include some of the most overspecialized types of that group. Succeeding coals show almost none of these forms.

In West Virginia the Upper Freeport coal is separated from the Lower Freeport coal by a conspicuous floristic change. The several Kittanning coals are similar to the Lower Freeport. A major break comes also in the Monongahela series at the thick Benwood and Arnoldsburg limestones, for the Uniontown-Waynesburg flora is quite unrelated to the general type of megaspore flora found in the Little Pittsburgh, Pittsburgh, Redstone, and Sewickley coals. A study of the Arnoldsburg and/or Lower Uniontown coals has not been made, so the actual time of this change cannot be delimited.

Such major floristic changes should be considered in dividing the Pennsylvanian

rock sequence of the Appalachian Basin. Considerable additional evidence from further studies is required to substantiate some breaks, but the microfossils, along with other available information, can be put to use at any time.

REMARKS

The writer is keenly aware of the possibility of using microfossil floras in the correlation of coal seams, for the determination of coal reserves, for the recognition of split seams and erosional or nondepositional unconformities, and as a means of reducing the expense of prospecting and proving an area by the usual methods. But it is a painful fact that the whole science of micropaleobotany has failed to meet its challenge, except for a few instances of local application of its principles. This failure, which need be only temporary, is principally attributable to the confused status of microfossil nomenclature and identification. As already stated, the present confusion stems from the difficulty of identification of material, from the uncertain status of many species, and, basically, from the relative difficulty of determining the actual specific value of the variations between one microfossil and another.

Many species have been created that are doubtless indistinguishable from other species described. Inadequate illustrations and incomplete or muddled descriptions occur commonly in the literature. It should again be recalled that the present work is *not* a taxonomic treatise, for in most cases the illustrations would be quite as inadequate as some already referred to in the literature of this field, and there is no intention of giving here the detailed scientific descriptions which will be necessary when a complete study is published. The use of a common, simple terminology by all workers, regardless of nationality, must be forthcoming if the science is to advance or ever be of economic value. The confused status at present must gradually be resolved by a systematic comparison of the type material of all conflicting and intergrading species. Description of species by the use of symbols or formulas is decried as an additional complication. The paper and space saved are relatively unimportant, and the time saving is questionable. Publication of descriptions using such abbreviations or other laboratory devices could be likened to the publication of the untranscribed shorthand notes of a secretary rather than to the intended word message. It is probable that systems can be worked out to use such short cuts in the individual laboratories, but a straightforward word description accompanied by adequate photographs and/or diagrammatic drawings will be most useful to all.

It is not deemed reasonable that any commercial laboratory be required to publish, or censured for not publishing, systematic works. The arbitrary numerical designations which have appeared in the literature are justifiable for the average laboratory technician to use, especially under the present status of taxonomic confusion. If taxonomic uncertainty is reduced or eliminated, most laboratories will favor using a binomial system of nomenclature established on biological principles.

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PALEOZOIC CORDILLERAN GEOSYNCLINE AND RELATED OROGENY

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ABSTRACT

The Cordilleran geosyncline in Paleozoic time consisted of two main troughs—a western, the Pacific, and an eastern, the Rocky Mountain. The sediments of the western trough from California to Alaska are characterized by a large amount of volcanic material and graywacke in every system. Phyllites, slates, argillites, schists, gneisses, recrystallized chert, marble, metaconglomerate, meta-andesite, and various metamorphosed pyroclastics make up the thick sequences. Great batholiths of Mesozoic age invaded the sediments of the Pacific trough; but in only one place, the Idaho batholith, did they reach the Rocky Mountain trough.

Dynamic metamorphism and the great intrusives, together with blanketing Cenozoic deposits, have left the medial area between troughs little known. The Rocky Mountain trough contains mostly marine limestones, shales, and sandstones. Many of the sands are cemented with silica and termed "quartzite," but little metamorphism has occurred, and none has been assigned to Paleozoic orogeny.

The volcanic materials are nearly all andesites and came from the west. Since the sites of extensive accumulation of andesitic flows and debris today are in troughs adjacent to island archipelagos, it is suggested that a volcanic archipelago flanked the Pacific trough on the west. The Paleozoic graywackes and conglomerates also came from the west and from Paleozoic rocks already metamorphosed before erosion. Several angular unconformities indicate orogeny in the belt to the west. The island archipelago was therefore a site of continuing orogeny during the Paleozoic. It was very similar to the island arcs of the present Japanese archipelago.

Instead of a rigid land of continental proportions, with a shore in western Montana, a volcanic orogenic archipelago is believed to have existed, which for the most part lay west of our modern coast.

PALEOZOIC CORDILLERAN GEOSYNCLINE

DIVISIONS AND THEIR CHARACTERISTICS

Charles Schuchert[1] is probably more responsible than anyone else for the use of the expression "Cordilleran geosyncline" in describing the basins of accumulation of sediments along the western part of the continent. He also used the term "Rocky Mountain geosyncline." The sediments of the seaways that shifted over the entire western third of the continent from the Dakotas to California were included. During the Mesozoic his "Cordilleran intermontane geanticline" split the over-all broad and irregular basins into two "geosynclines"—a western and an eastern. He emphasized the changes in the shorelines of the seaways, whereas in the following pages the troughs of maximum subsidence and the sediments in them will receive special attention.

In Paleozoic time the Cordilleran geosyncline consisted in a very general way of two main troughs, a western, here called the "Pacific," and an eastern, the Rocky Mountain (see Figs. 1 and 2). The medial area was also one of heavy sedimentation, but not much is known about it, especially in Canada.

[1] *Historical Geology* (several eds.; New York: John Wiley & Sons).

Linlithgow Library.
Imperial Agricultural Research Institute.

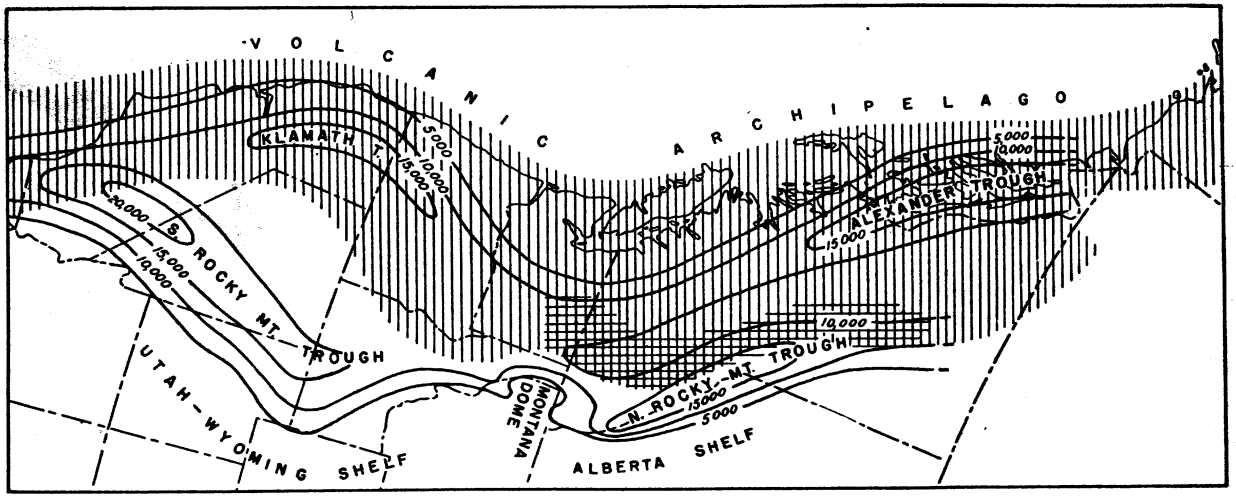


FIG. 1.—Pre-Devonian Paleozoic troughs of the Cordilleran geosyncline. The horizontally ruled area represents the distribution of the volcanic archipelago assemblage of sediments. The cross-ruled area is the Shuswap terrane.

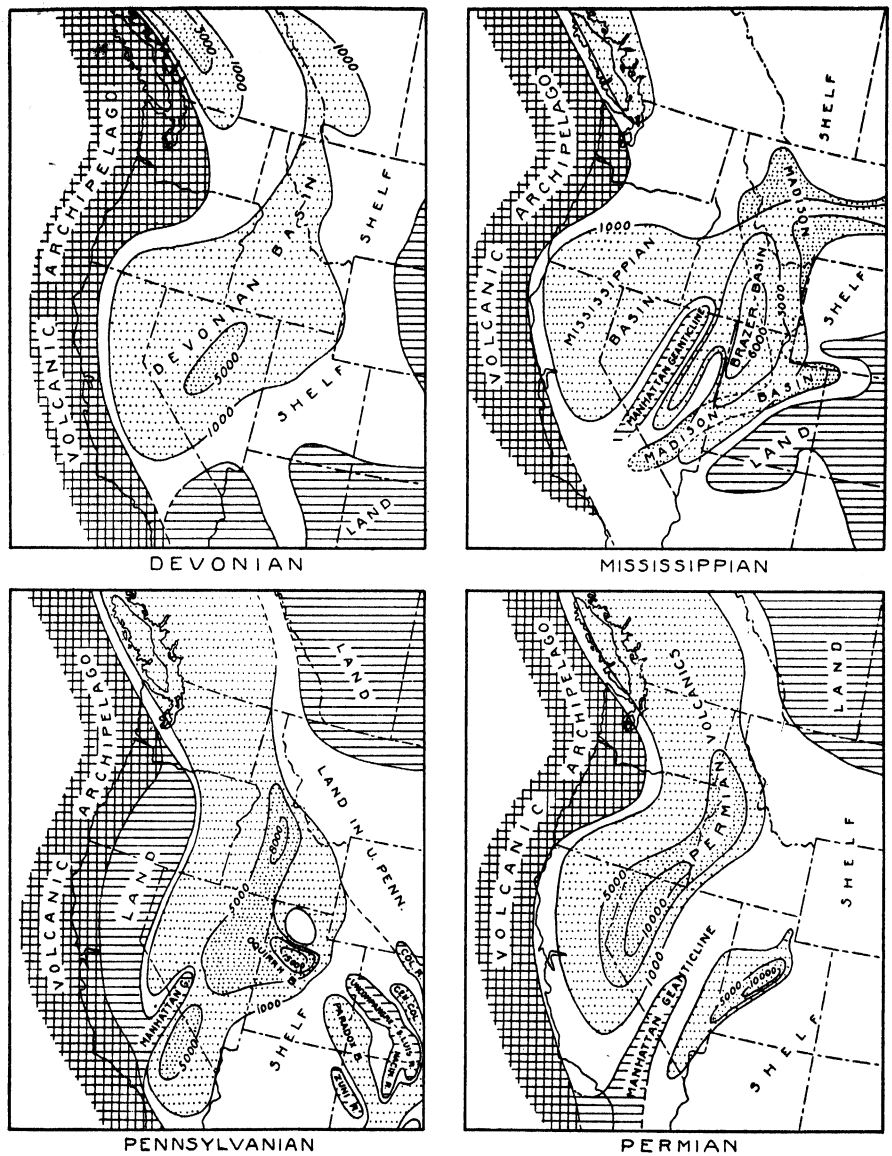


FIG. 2.—Paleotectonic maps of the Cordilleran region in late Paleozoic time. Cross-ruled area is the volcanic archipelago and orogenic belt. Horizontally ruled areas were uplifted and eroded during the period designated. White areas sank and were covered with less than 1,000 feet of sediments (after consolidation). Lightly stippled areas sank sufficiently to be covered with more than 1,000 feet of sediments. Denser stippling denotes areas that sank more than 5,000 feet, except on the Mississippian map, where it distinguishes the Madison (Lower Mississippian) from the Brazer (Upper Mississippian). The "lands" of each period are areas that actually rose and are not those simply that remained above sea level as a result of a previous uplift.

The western trough probably sank more and received a greater thickness of sediments than did the eastern, but the extent of sediments in both was of geosynclinal proportions. The greatest difference lies in the character of the sediments. The Pacific trough received a dominant amount of volcanic material and graywacke, whereas the Rocky Mountain trough was filled with sandstones, shales, and limestones. The volcanic material in the Pacific trough is in several forms—tuffs, flows, volcanic conglomerates, and various pyroclastics. The volcanics and graywackes occur in every stratigraphic system from Ordovician and older to Cretaceous. The Permian especially was a time of excessive volcanism, and the volcanics of that period have been traced from California and western Nevada to Alaska.[2] In the Humboldt Range of northwestern Nevada, over 10,000 feet of Permian strata, largely volcanic, have been identified.

Another characteristic of the sediments of the Pacific trough is their metamorphism. Phyllites; slates; argillites; quartz, chlorite, hornblende, and calcareous schists; hornblende gneiss; recrystallized chert; marble; metaconglomerate; meta-andesite and various metamorphosed pyroclastics make up the thick sequences. Still another characteristic is the presence of great intrusive bodies of a later age and the metamorphism of the sediments about the intrusions.

The sediments of the Rocky Mountain trough, on the other hand, are not metamorphosed. Many of the sands are cemented with silica and termed "quartzite," but little dynamic metamorphism incident to Paleozoic, Mesozoic, or Tertiary orogeny has occurred.

[2] H. E. Wheeler, "Permian Volcanism in Western North America," *Proc. 6th Pacific Sci. Cong.*, Vol. I (1939), pp. 369–76.

ROCKY MOUNTAIN TROUGH

Pre-Devonian basins.—The Rocky Mountain trough is noted for its Cambrian sections. At one locality in southern Nevada and California, 17,000 feet of Lower, Middle, and Upper Cambrian strata have been measured. This thickness gradually decreases northward through Utah to northern Idaho as the three divisions of the Cambrian successively overlap one another. A small positive area in northern Idaho and southernmost British Columbia separates the southern Cambrian trough from the northern. The positive area has been called the "Montana Island"[3] and "Montania."[4] In order to conform to a nomenclature adopted for the book of which this article is to be a chapter, the uplift will be called a "dome." It is the present site of exposure of the great Beltian series of Proterozoic age.

North of the dome a second deep trough existed, in which the three divisions of the Cambrian also occur. The total thickness of the Cambrian sediments there is about 17,000 feet. It is in these strata that Walcott made his famous studies of Cambrian life. Out of the strata are sculptured the high and scenic Canadian Rockies of the Banff and Jasper regions.

The Ordovician strata are fairly thick in both the southern (5,000 feet) and the northern (2,500 feet) divisions of the trough, but Silurian strata occur principally in the southern.

The distribution of the early Paleozoic strata is fairly well known westward to

[3] Schuchert, 1924 ed. of ftn. 1.

[4] C. F. Deiss, "Cambrian Geography and Sedimentation in the Central Cordilleran Region," *Bull. Geol. Soc. Amer.*, Vol. LII (1941), pp. 1085–1116.

central Nevada in the southern trough,[5] but the lithology of the connecting link with the formations of the Pacific trough is largely unknown. This is due to the great Jurassic batholiths and the cover of Tertiary volcanics (examine cross sections of Fig. 3).

Although a trough in the site of the Canadian Rockies extending northward to the Mackenzie delta is postulated by several students of paleogeography, the western flank of the trough has not been located. C. D. Walcott's[6] sections in southern British Columbia were isopached for each division of the Cambrian and for the Ordovician, and then the four isopach maps were superposed and collectively contoured. The result showed a progressively deepening trough westward, with only the slightest indication of a trough axis before metamorphic rocks appear, which have been called "pre-Cambrian" (see Fig. 1). It is now fairly well demonstrated that the metamorphic rocks which spread to the Pacific trough are not only of pre-Cambrian age but also of Paleozoic age, and therefore it seems likely that sedimentation occurred from the northern Rocky Mountain trough to the Pacific trough. A slight swell may have separated the two; but, if it existed at all, it should be considered as a linear zone of the crust that failed to subside as rapidly as the troughs on each side.

Far north along the Mackenzie River in the Northwest Territories, 3,500–4,500 feet of pre-Devonian Paleozoic strata have been observed,[7] which are chiefly Silurian and Ordovician in age. They are in part an evaporite sequence. So little is known of their distribution that no attempt has been made to extend the isopachs of the northern Rocky Mountain trough to this region.

Devonian basin.—Although Devonian strata are found nearly everywhere west of the transcontinental arch, they are over 1,000 feet thick only in the western part of the general Rocky Mountain area (see Devonian map of Fig. 2). In the Roberts Range, Nevada, C. W. Merriam[8] has described 4,465 feet of Devonian beds, and at near-by Eureka he has found 4,000–5,000 feet of them. They are composed chiefly of limestones and dolomites, their fossil content indicates a rather complete section, and the broad trough in which they accumulated subsided during most of Devonian time.

Another fairly deep Devonian basin is known in the far north—the Norman Wells district on the Mackenzie River in the Northwest Territories.[9] A maximum thickness of 3,500 feet of Middle and Upper Devonian beds has been described; but, since they are at the surface over an extensive area northward to the Mackenzie delta or are overlapped southward by the Cretaceous, it appears that the original thickness may have been greater and was comparable to the thickness in the Nevada basin.

Mississippian basins.—The Lower Mississippian Basin, in which the Madison limestone and equivalents were deposited, is somewhat different from the

[5] T. B. Nolan, "The Basin and Range Province in Utah, Nevada, and California," *U.S. Geol. Surv. Prof. Paper 197-D* (1943), pp. 141–96.

[6] "Pre-Devonian Paleozoic Formations of the Cordilleran Provinces of Canada," *Smithsonian Misc. Coll.*, Vol. LXXV (1928), pp. 175–368.

[7] J. S. Stewart, "Petroleum Possibilities in Mackenzie River Valley, N.W.T.," *Trans. Can. Inst. Min. Met.*, Vol. XLVII (1944), pp. 152–71.

[8] "Devonian Stratigraphy and Paleontology of the Roberts Mountains Region, Nevada," *Geol. Soc. Amer. Special Paper 25* (1940), pp. 1–144.

[9] See ftn. 7.

Upper Mississippian Basin, in which the Brazer formation and equivalents accumulated, and therefore the Mississippian crustal movements will be treated in two stages (see Mississippian map of Fig. 2). The Madison Basin, as defined by the 1,000-foot-thickness contour, is limited to a gangling trough in southern Nevada, central Utah, eastern Idaho, and central Montana. The thickness of sediments is greatest near Pioche, Nevada, where L. G. Westgate and A. Knopf[10] record about 3,000 feet. They make up the Bristol Pass limestone and the Peers Spring formation. At Three Forks, Montana, 1,800 feet of beds have been measured[11] and, farther north in the Saypo quadrangle, 1,000 feet.[12] The suggested thinning northward toward the International Boundary is substantiated by the presence of nothing but Beltian (Proterozoic) rocks in the Glacier National Park area.

J. S. Williams[13] believes a landmass must have existed in northern Colorado and south-central Wyoming from which sandstones and various red beds were derived.

In the Death Valley region of California the Lower Mississippian beds are probably over 2,100 feet thick, and therefore it appears that the Madison trough extended southwestward through the Pioche district, where the Lower Mississippian is about 2,800 feet thick, to Death Valley. No thick Upper Mississippian section is known in the area of southern Nevada and northwestern Arizona.

Larger and deeper basins formed in Upper Mississippian time than in Lower. The Brazer Basin of northern Utah and southeastern Idaho is well known. J. S. Williams[14] has measured nearly 3,000 feet in Big Cottonwood Canyon, 1,750 feet in Weber Canyon, and 3,100 feet in Wellsville Mountain (all in the Wasatch Mountains); and G. R. Mansfield[15] lists over 1,100 feet in southeastern Idaho. In the Makay region of central Idaho, Umpleby *et al.*[16] state that the section may be much more than 6,000 feet thick. This great thickness corresponds with Nolan's 6,000 feet for the Woodman and Ochre Mountain formations of Gold Hill, Utah.

At Pioche, Nevada, the Upper Mississippian is represented probably by the Scotty Wash quartzite and by the lower half of the Bailey Spring limestone, and these beds are about 2,000 feet thick.

In the San Francisco District, Utah, the Topache limestone is probably Upper Mississippian and is 1,500 feet thick. No Lower Mississippian exists there.

The Lower Mississippian strata of Arizona do not exceed 300 feet in thickness.[17] Upper Mississippian are thin and occur only in the southeastern corner.

In the Muddy Mountains, Nevada,

[10] "Geology and Ore Deposits of the Pioche District, Nevada," *U.S. Geol. Surv. Prof. Paper 171* (1932).

[11] George W. Berry, "Stratigraphy and Structure at Three Forks, Montana," *Bull. Geol. Soc. Amer.*, Vol. LIV (1943), pp. 1–30.

[12] C. F. Deiss, "Stratigraphy and Structure of Southwest Saypo Quadrangle, Montana," *Bull. Geol. Soc. Amer.*, Vol. LIV (1943), pp. 205–62.

[13] "Carboniferous Formations of the Uinta and Northern Wasatch Mountains, Utah," *Bull. Geol. Soc. Amer.*, Vol. LIV (1943), pp. 591–624.

[14] *Ibid.*

[15] "Geography, Geology, and Mineral Resources of Part of Southeastern Idaho," *U.S. Geol. Surv. Prof. Paper 152* (1927), pp. 1–409.

[16] J. B. Umpleby, L. G. Westgate, and C. P. Ross, "Geology and Ore Deposits of the Wood River Region, Idaho," *U.S. Geol. Surv., Bull. 814* (1930).

[17] A. A. Stoyanow, "Correlation of Arizona Paleozoic Formations," *Bull. Geol. Soc. Amer.*, Vol. XLVII (1936), pp. 459–540; "Paleozoic Paleogeography of Arizona," *ibid.*, Vol. LIII (1942), pp. 1255–82.

Longwell estimates about 600 feet of Lower Mississippian and perhaps over 900 feet of Upper Mississippian.

In Upper Mississippian time a fairly pronounced basin developed in east-central Montana and western North Dakota, and in it the Big Snowy group was deposited. The beds attain a maximum thickness of about 1,500 feet.[18] It is not clear whether the sediments for the Big Snowy Basin came from a highland on the south or the north.

Manhattan geanticline.—Toward the end of the Devonian period, according to Nolan,[19] a geanticline began to rise in west-central Nevada, approximately along the line of maximum sedimentation of the early Paleozoic (see Figs. 2 and 3). The uplift divided the geosyncline into a western and an eastern trough, and the distribution of Devonian sediments is reflected in two ways, viz., by the almost complete removal of the earlier Devonian deposits along the axis of the arch and by an eastward shift to the vicinity of Eureka, Nevada, of the zone of maximum sedimentation. Since most other major structural features have been given names, it is desirable for the purpose of illustration on the tectonic maps to assign one to the geanticline in central Nevada. The axis of it runs approximately through Manhattan, where a good section has been measured; hence the name "Manhattan geanticline" seems appropriate.

The large proportion of clastics in the Mississippian and in the lower part of the Pennsylvanian sequences in California and the coarse material in the upper part of the Pennsylvanian at Eureka, Nevada, indicate that elevation of the geanticline continued and that it was subject to erosion throughout nearly all Mississippian and Pennsylvanian time.[20]

The Permian marked the close of the elevation and planation. Slight submergence probably permitted deposits of that period to be laid down over the entire area.

The elevation of the arch was accomplished by broad warping, but folding occurred locally, so that angular discordances up to 35° exist, such as in the Candelaria Hills–Miller Mountain area, where the underlying Ordovician strata are truncated by the Permian.

The axis of the geanticline has been traced as follows:[21] Near Austin, Nevada, Pennsylvanian limestone is reported to rest on slates and limy shales of probable Ordovician age. To the south the axis is drawn west of the Inyo Range because of the similarity of the section to that of Eureka and its difference from the Carboniferous Calaveras formation in California. What happened to the geanticline southward is unknown. It may have merged with a broad area of uplift or orogeny. The course of the geanticline west of the Inyo Range marks the site of greatest sediment accumulation in pre-Carboniferous time.

Still farther north the section in the Ruby Mountains[22] indicates that the axis of the upwarp is to the west. It is, therefore, headed toward the Bayhorse and Wood River regions of Idaho, but the thick sections there are typical of those of the undeformed basin, and hence the geanticline appears to have died out approximately at the Idaho line.

[18] H. W. Scott, "Some Carboniferous Stratigraphy in Montana and Northwestern Wyoming," *Jour. Geol.*, Vol. XLIII (1935), pp. 11–32; N. Ballard, "Regional Geology of Dakota Basin," *Bull. Amer. Assoc. Pet. Geol.*, Vol. XXVI (1942), pp. 1557–84.

[19] See ftn. 5.

[20] *Ibid.*

[21] *Ibid.*

[22] R. P. Sharp, "Stratigraphy and Structure of the Southern Ruby Mountains, Nevada," *Bull. Geol. Soc. Amer.*, Vol. LIII (1942), pp. 647–90.

Pennsylvanian and Permian basins.—Pennsylvanian sediments are widespread in the western United States, and in places they attain great thicknesses. According to Nolan,[23] they show more abrupt changes in thickness within relatively short distances than do the Upper Mississippian sections; and this variability, together with the higher proportion of clastic material in the beds and the presence of local unconformities, indicates considerable instability of the broad geosynclinal area during Pennsylvanian time. The greater clastic content appears to be especially characteristic of the sedimentary sections near the eastern and southeastern borders of the province and suggests the presence in this region of marked crustal activity. The Ancestral Rockies may have furnished much of the sand (see Pennsylvanian map of Fig. 2).

The basin sank most in west-central Utah, where 18,000 feet of beds are recorded.[24] At several other near-by places 8,000–10,000 feet have been measured. The deep basin is approximately contoured on the map and is labeled the "Oquirrh Basin," from the Oquirrh Range, where the greatest thickness occurs.

Another fairly deep basin formed in southern Nevada and southern California, where a maximum of 7,750 feet of Pennsylvanian rocks are known.[25]

Pennsylvanian beds in western Nevada, northern California, Oregon, and Washington are dominantly volcanic and will be described under the next heading.

[23] See ftn. 5.

[24] James Gilluly, "Geology and Ore Deposits of the Stockton and Fairfield Quadrangles, Utah," *U.S. Geol. Surv. Prof. Paper 173* (1932).

[25] Adolph Knopf, "A Geologic Reconnaissance of the Inyo Range and the Eastern Slope of the Southern Sierra Nevada, California, with a Section on the Stratigraphy of the Inyo Range by Edwin Kirk," *U.S. Geol. Surv. Prof. Paper 110* (1918).

East of the Manhattan geanticline the strata are sandstones, shales, and limestones.

Permian sediments are very widespread in the western Cordillera and consist, in general, of two divisions: the eastern marine sequences and the western bedded volcanic series. The eastern trough is not large if marked off by the 1,000-foot isopach (see Permian map, Fig. 2); but in one place in central Utah it sank over 10,000 feet and was filled with marine limestone, shale, and sandstone beds. Over shallow shelves to the north, especially in northern Utah, eastern Idaho, western Wyoming, and south-central Montana, great quantities of chert and phosphatic limestone and shale were deposited.

PACIFIC TROUGH

In the Klamath Mountains.—The pre-Devonian rocks in the Klamaths are metamorphosed and, their age is a matter of guess. The section represented (Figs. 3 and 4) is that given by N. E. A. Hinds[26] for the southern Klamath Mountains in northern California, specifically in the Redding, Weaverville, and northwestern part of the Red Bluff quadrangles. Formations dated because of fossil content are of Devonian, Mississippian, and Permian age. Formations older than Middle Devonian, but in which no fossils have been found, are the Copley meta-andesite, the Chanchelulla formation, the Salmon schist, and the Abrams schist. The two schists have been called collectively the "Siskiyou terrane." Although the schists are separated from each other by an unconformity and are, therefore, of different ages, they form a distinct unit because they are equally

[26] "Paleozoic Section in the Southern Klamath Mountains, California," *Proc. 6th Pacific Sci. Cong.*, Vol. I (1939), pp. 273–88.

metamorphosed and much more so than the later formations of the area. They accumulated to a thickness of at least 5,000 feet in a trough that extended for a known distance of 200 miles north and south in northern California and southwestern Oregon. Pebbles and boulders of the Salmon and Abrams schists are found in conglomerates of the overlying Chanchelulla formation. Metamorphism, deformation, and deep erosion of the schists preceded the accumulation of the Chanchelulla, and therefore it is possible that they are much older. Hinds[27] believes that they are both pre-Silurian. In cross section *B–B′* of Figure 3 the Salmon schist is questionably shown to be Cambrian and the Abrams pre-Cambrian. The Salmon is composed of hornblende and chlorite schist and was originally a series mostly of fine, water-laid, basic volcanic-ash beds. Since most other Paleozoic formations in the Pacific trough contain much volcanic material, it seemed logical in drawing the cross section to correlate the Salmon with the Cambrian. It is purely an assumption, however.

The Abrams schist is separated from the Salmon by an erosional unconformity[28] and is made up dominantly of mica schist, with mica quartz schist and metaquartzite present in places. It is the oldest rock known in the Klamath Mountains and may be pre-Cambrian, but a Cambrian age is certainly not excluded. It is estimated to be at least 5,000 feet thick.

The Chanchelulla formation overlies the schists "with moderate angular and distinct erosional unconformity"[29] and forms a broad belt which, as far as is known, runs in a northwest direction. The base is composed largely of gray, thinly bedded, recrystallized chert. Interbedded are micaceous, graphitic, and calcareous schists, quartzite, metaconglomerate, and marble. Toward the top of the formation the proportion of mica schist, partially metamorphosed sandstone, conglomerate, and limestone is considerably greater. The strata have been strongly folded and moderately metamorphosed, though recrystallization is much less intense than in the underlying schists. The formation exceeds 5,000 feet in thickness.

After the sediments were deposited,

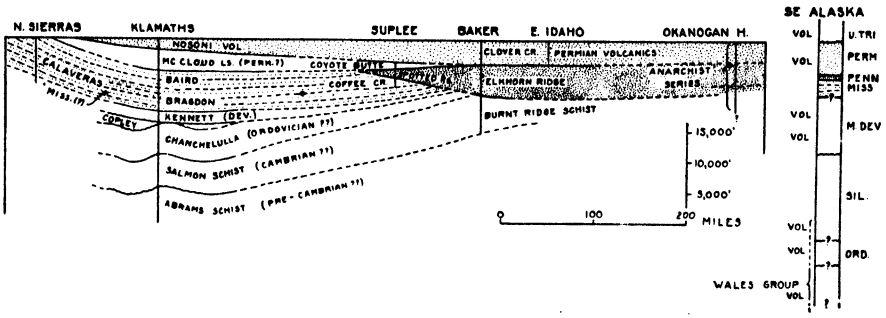


FIG. 4.—Possible correlation of the volcanics and metamorphic sequences of the Cordilleran geosyncline in northern California, Oregon, and Washington. The vertical lines indicate the extent of the exposed sections. The drawing is an attempt to restore the relations as they were before the post-Permian, pre-Triassic folding. "Vol" under "SE Alaska" indicates position of various volcanics in the systems.

[27] *Ibid.*

[28] *Ibid.*

[29] *Ibid.*

they were invaded by sills of andesite and andesite prophyry, some 200–500 feet thick, with great areal extent. The andesites have since been metamorphosed to greenstone or chlorite schist. The meta-andesites closely resemble the overlying Copley volcanics, and their degree of metamorphism is similar. The suggestion is strong, therefore, that the basic rocks of the Chanchelulla formation are intrusive components of the Copley eruptives.

The Chanchelulla is pre-Copley, which is pre-Middle Devonian. Hinds believes the Chanchelulla is early Paleozoic, and it seems to correlate on the cross section conveniently with the Ordovician system farther east.

Widely distributed through the Klamath and Siskiyou Mountains are remnants of a great volcanic series called the "Copley meta-andesite." The formation consists of coarse and fine andesitic and basaltic pyroclastics, with interbedded flows of pyroxene, andesite, and basalt. Sections 1,500 feet thick have been measured, but good sections are hard to find, and the original thickness is not known.

The Copley volcanics have generally been metamorphosed to greenstone schists, but the intensity of recrystallization varies from place to place. Porphyritic flows, both vesicular and amygdaloidal, and brecciated, blocky, or clinkery lavas are not uncommon. Near the contacts of the Nevadan (late Jurassic) intrusions, the volcanics have been recrystallized generally to chlorite schist. The metamorphism by the granite intrusives has been superimposed on the earlier metamorphism.

The Copley volcanics lie with "pronounced angular and erosional discordance" upon the Siskiyou schists and the Chanchelulla formation and are overlain unconformably by the Middle Devonian Kennett formation. The Copley is shown related to the Silurian strata of the Taylorsville region, but this is simply a matter of necessity after the Chanchelulla was made equivalent to the Ordovician. J. S. Diller[30] has collected Silurian fossils from the oldest rocks found in the Taylorsville district, but these rocks are chiefly thin-bedded quartzite with dark slaty shale and limestone and do not resemble the Copley volcanics.

In Devonian time the western margin of the broad basin in Nevada was filled chiefly with black clays, now shales and slates. They are highly siliceous and contain thin beds of sandstone and lenses of fossiliferous limestone, now largely recrystallized. The Devonian beds are known as the "Kennett formation" and crop out in two rather restricted belts in the Klamath Mountains. No strata yet identified as Devonian are known in California south of the Klamaths in the Sierra Nevada Mountains or the Coast Ranges, and none have been noted in Oregon to the north.

The Kennett lies unconformably on the deformed and eroded Copley meta-andesite, and the basal strata of the Kennett contains rock fragments, evidently from the volcanics.[31]

The Devonian strata were originally thicker than now, having been subjected to erosion and entirely removed in places. As a consequence, the Mississippian beds rest directly on the Copley volcanics over large areas.

The Mississippian is made up of two formations in the Klamath Mountains—the Bragdon and the Baird. They are probably the most widespread formations in the region. The Bragdon is chief-

[30] "Geology of the Taylorsville Region, California," *U.S. Geol. Surv. Bull. 353* (190).

[31] See ftn. 26.

ly shale and slate, generally gray, in contrast to the black shale and slate of the older Kennett formation. Some sandstones are conglomeratic near the base and contain fragments of both the Kennett and the Copley formations. Within the Redding quadrangle a volcanic sequence, called the "Bass Mountain basalt," is present. The Bragdon may exceed 6,000 feet in thickness in places. The Bass Mountain volcanic sequence contains many tuff beds in places. Its position on Bass Mountain, according to Hinds, is in the lower part of the Bragdon formation.

The Baird formation consists largely of sandstone and tuff, but the upper part has calcareous and siliceous slates. It is about 700 feet thick and apparently rests conformably on the Bragdon.[32]

The McCloud limestone is a highly fossiliferous formation and was probably a massive cherty limestone, but now, because of the Jurassic intrusions, it is mostly metamorphosed in various degrees to marble. It reaches a maximum thickness of 2,000 feet. Its fossils were first thought to represent a Pennsylvanian age, but a recent study by Wheeler[33] shows them to be Lower Permian.

The McCloud limestone overlies the Mississippian Baird formation disconformably; hence it appears that most of the Pennsylvanian was a time of emergence.

The Nosoni formation is composed of basaltic agglomerates; lithic crystal tuffs; flows of andesite and of olivine basalt; dark brown, fossiliferous, shaly limestone; and dark-gray to brown tuffaceous shales and slates. The maximum thickness measured by Hinds is 1,200 feet. It is considered to be upper Lower Permian.[34] Hinds[35] believes that the Nosoni volcanics rest on an erosion surface on the McCloud limestone.

In the Sierra Nevada Mountains.—In the northern Sierra Nevadas a metamorphic formation of Carboniferous age is widespread and consists chiefly of black phyllite, with subordinate fine-grained quartzite, limestone, and chert. It is known as the "Calaveras formation." Associated and in part interbedded with the formation are green amphibolite schists of contemporaneous age. From fossils, found chiefly in the limestone, the Calaveras formation is known to be, in part at least, of Carboniferous age, but parts of it as mapped may be Devonian and Triassic. Because of the metamorphosed condition of the rocks in which the fossils are found, it has been difficult for paleontologists to determine to what part of the Carboniferous the faunas belong. Girty[36] thought groups of Calaveras fossils from the Taylorsville region are more closely related to the Baird, now recognized as Mississippian, than to the McCloud limestone, now believed to be Permian. The amphibolite schists were originally fine pyroclastics.[37]

The bedded rocks are most abundant in the northern Sierras, but southward they become increasingly metamorphosed, and progressively greater areas are occupied by granitic intrusives. In the Tehachapi Mountains and the southern Coast Ranges, pregranitic rocks are present, but highly altered, and their ages are speculative.

[32] *Ibid.*

[33] *Ibid.*

[34] H. E. Wheeler, "Fusulinids of the McCloud and Nosoni Formations of North California," *Bull. Geog. Soc. Amer.*, Vol. XLIV (1933), p. 218.

[35] See ftn. 26.

[36] See ftn. 30.

[37] Adolph Knopf, "The Mother Lode System of California," *U.S. Geol. Surv. Prof. Paper 157* (1929).

In central Oregon.—A heterogeneous group of east-west-trending ranges and dissected lava plateaus, known collectively as the "Blue Mountains uplift" or the "Blue Mountains–Ochoco Mountains uplift,"[38] extends from central to eastern Oregon. The ranges are formed of Paleozoic and Mesozoic sediments and lavas and Mesozoic plutons, and the complex protrudes island-fashion through the Columbia River lava fields. The oldest beds that crop out are Lower Carboniferous limestones and calcareous sandstones[39] (see Figs. 4 and 5). About 1,000 feet of them are exposed, and they are called the "Coffee Creek formation." No volcanic materials have been noted.

Overlying the Coffee Creek formation is the Spotted Ridge formation. The exact contact relations have not been observed, but, if an unconformity does exist, it is probably not angular and does not represent much of a time break. The Spotted Ridge consists of plant-bearing sandstones and mudstones; conglomerates containing diorite, andesite, and dacite boulders; and bedded chert. It may be 1,500 feet thick. The plants are believed to be Lower Pennsylvanian.

The Coffee Creek and Spotted Ridge formations are reported as intensely folded, but no mention is made of metamorphism.[40] They lie in a tectonic belt of deformed strata in which the rocks on the west (Klamaths) and on the east (Baker area) are metamorphosed, and it is puzzling that they also are not metamorphosed.

[38] A. C. Waters, "Summary of the Sedimentary, Tectonic, Igneous, and Metalliferous Geology of Washington and Oregon, Ore Deposits of the Western United States," *Lindgen Volume, A.I.M.E.* (1933), pp. 253–65.

[39] C. W. Merriam and S. A. Berthiaume, "Late Paleozoic of Central Oregon," *Bull. Geol. Soc. Amer.*, Vol. LIV (1943), pp. 145–72.

[40] *Ibid.*

The Spotted Ridge is overlain by the Coyote Butte formation. A slight angular unconformity separates the two. The Coyote Butte is composed almost entirely of massive limestones. Some chert pebble conglomerates are present near the base. The age is probably Lower Permian.

A prominent angular unconformity exists between the Paleozoic beds of central Oregon and the overlying Triassic conglomerates, which attain a thickness of 4,000 feet.

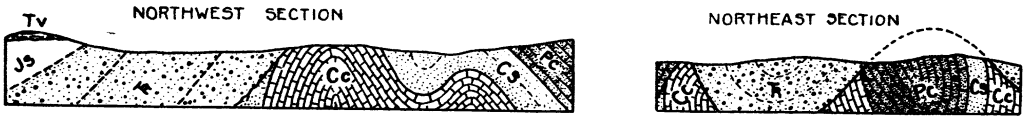
In eastern Oregon.—In the Baker quadrangle of eastern Oregon, James Gilluly[41] described a formation—the Burnt River schist—which, chiefly because of greater metamorphism than that of known Carboniferous rocks near by, he cautiously treats as older. The rock varieties are greenstone schist, quartz schist, conglomerate schist, limestone, slate, and quartzite, and they make up a series at least 5,000 feet thick, perhaps several times as much. The various types mentioned grade into one another.

Gilluly visualizes the origin of the strata as follows:

> . . . . pyroclastic material was added in amounts varying from time to time to a basin of sedimentation to which at some times sand and at others clay, with some carbonates, were being supplied. When volcanic contributions were small the deposits were such as have yielded the quartzites and carbonaceous slates now found, but when the volcanic material increased relative to the normal terrigenous sediment the deposits were such as have yielded the intermediate rocks. . . . . At times such floods of volcanic material were contributed that practically unmixed tuff was formed.

The Burnt River schist has lithologic similarities with the Calaveras formation but differs, it seems, in generally having

[41] "Geology and Mineral Resources of the Baker Quadrangle, Oregon," *U.S. Geol. Surv. Bull. 879* (1937).



GRINDSTONE-TWELVEMILE CREEKS AREA, CENTRAL OREGON (MERRIAM AND BERTHIAUME)

Cc, Coffee Creek fm. (Lower Carb.) Cs, Spotted Ridge fm (Penn.) Pc, Coyote Butte fm. (Perm.)

1 MILE

NORTH-SOUTH SECTION NEAR BAKER, OREGON (GILLULY)

brs, Burnt River schist; Ce, Elkhorn Ridge argillite (Penn.?) Ccg, Clover Creek greenstone (Perm.)
qd, biotite-quartz diorite; sg, silicified gabbro; mg, metagabbro; gb, gabbro

5 MILES

FIG. 5.—Cross sections in central and eastern Oregon

greater metamorphism and an absence of chert. The Burnt River appears from published descriptions to be surprisingly similar to the Salmon schist of the Klamaths, which is probably pre-Silurian (see Figs. 4 and 5).

Above the Burnt River schist is the Elkhorn Ridge argillite, about 5,000 feet thick. It is probably the most widespread of the pre-Tertiary formations and is a thick series of argillite, tuff, and chert with subordinate limestone and greenstone masses. A number of large intrusive bodies have been noted in the east-west belt of argillite; and these, together with the overlapping Cenozoic rocks, effectively prevent the recognition of contacts and the determination of stratigraphic relationships. The beds are Pennsylvanian in age because of fusulinid fossils found in the limestones. Beds younger than Pennsylvanian may have been included in the formation as mapped.[42]

The whole formation is provisionally considered marine. The tuffaceous argillite, the tuff, and the tuffaceous limestone all clearly attest notable pyroclastic contributions to the formation, and it is highly probable that cherts as numerous and thick as those in this formation may be considered evidence of igneous contribution also.

The association of limestone with volcanic materials may have no genetic significance, but a dependency is suspected because volcanism might have raised the temperature of the sea and hence decreased the solubility of the lime.[43]

The Clover Creek greenstone overlies the Elkridge argillite and consists of altered volcanic flows and pyroclastic rocks, with subordinate conglomerate, limestone, chert, and argillite. It is known to extend as far eastward as the Snake River Canyon and is therefore probably the same as the "Permian volcanics" of several areas in eastern Idaho. It is at least 4,000 feet thick,[44]

The effusive rocks in order of abundance are quartz keratophyre (lava-bearing albite), quartz keratophyre tuff, and meta-andesite. Fossils collected from the formation betray a Permian age.

The marine limestone and associated fossiliferous tuffs demonstrate a marine origin for at least part of the formation. The type of albitization which most of the volcanic rocks have undergone is common in demonstrably submarine volcanic rocks, and the association here with marine limestone suggests rather strongly that the Clover Creek greenstone is, in large part, of submarine origin.

In central Idaho and northwestern Nevada.—During Permian time the volcanic rocks of the Pacific trough spread eastward an unusual distance to central Idaho and northwestern Nevada. The sequence is 5,000 feet deep at Blairsden in the Sierra Nevadas and thickens eastward to 12,000 feet in the Humboldt Range, Nevada.[45] Northwestward into central Idaho it thins to about 4,000 feet (see map, Fig. 2).

In northern Washington and southern British Columbia.—Where the Okanogan River crosses the International Boundary, extensive areas of pre-Tertiary rocks are found. The pre-intrusive (pre-Jurassic) metamorphic rocks are called the "Anarchist series" and crop out in the Okanogan Range adjacent to the Okanogan Valley on the west and extensively in the Okanogan Highlands on the east.

[42] *Ibid.*

[43] *Ibid.*

[44] *Ibid.*

[45] See ftn. 5.

According to D. B. Krauskopf,[46] neither the top nor the bottom of the Anarchist series has been found, but at least 10,000 feet of beds exist. They can be divided rather vaguely into three divisions: the lower consists chiefly of gray to jet-black phyllites, with some interbedded quartzite and a little chlorite schist; the middle consists of limestone, massive quartzite, graywacke, conglomerate, some phyllite, and, to the north and east, much greenstone; the upper consists, for the most part, of greenstone, with some interbedded phyllite and quartzite. The albite in the greenstones of the upper division suggests a correlation with the keratophyres of eastern Oregon.

Regional metamorphism has converted the original sedimentary and volcanic rocks to a typical chlorite-zone assemblage. Near some of the plutonic bodies higher-grade contact metamorphism has been superimposed, yielding biotite and amphibolite schists and diopside rocks.[47]

A few fossils establish a marine origin for part of the series at least and a late Paleozoic age.

In Stevens County of northeastern Washington, C. E. Weaver[48] described the Stevens series, a group of metamorphic rocks with the great thickness, reportedly of 42,900 feet. It consists of quartzites, argillites, phyllites, dolomitic limestones, and schists. It is believed to be in part of Carboniferous age, but the lower parts are probably older. H. Bancroft[49] had previously found fragmentary plant fossils, which appeared to be Carboniferous.

The Carboniferous part of the Stevens series is probably equivalent to part of the Anarchist series on the west and to the Pend Oreille group[50] on the northeast along the forty-ninth parallel. The Pend Oreille is also considered to be in part Carboniferous. It and equivalents rest on the immensely thick Beltian strata of Proterozoic age which form a north-south belt in northern Idaho, western Montana, and British Columbia east of Kootenai Lake.

In the Okanogan Valley of British Columbia, C. E. Cairnes[51] recognizes a thick series of metamorphosed sediments, chiefly argillaceous—the Nisconlith series—and an overlying group of mixed volcanic and sedimentary rocks, which are fossiliferous in places and of late Carboniferous or Permian age—the Cache Creek group. The Nisconlith has been classed as both Paleozoic, Beltian, and pre-Beltian; but its age, other than pre–Cache Creek, is not known, because no fossils have been found in it.

Just north of the International Boundary in the Okanogan Valley (map 538*A*, Canada Dept. Mines and Resources) the Kobau group has been described. It is composed of much altered sediments and volcanic rocks. The sediments appear to have been originally argillaceous, and the amphibolitic and chloritic schists appear to have been andesitic lavas. Some Permian fossils have been found in limestone beds that appear to overlie the Kobau beds conformably, and hence the

[46] "Pre-Tertiary Intrusives of the Okanogan Valley (Washington) near the Forty-ninth Parallel," *Proc. 6th Pacific Sci. Cong.*, Vol. I (1939), pp. 223–29.

[47] *Ibid.*

[48] "The Mineral Resources of Stevens County," *Washington Geol. Surv. Bull. 20* (1920).

[49] "The Ore Deposits of Northeastern Washington," *U.S. Geol. Surv. Bull. 550* (1914), pp. 1–215.

[50] R. A. Daly, "Geology of the North American Cordillera at the 49th Parallel, Part II," *Can. Geol. Surv. Mem. No. 38* (Ottawa, 1912).

[51] "The Shuswap Rocks of Southern British Columbia," *Proc. 6th Pacific Sci. Cong.*, Vol. I (1939), pp. 259–72.

Kobau is regarded tentatively as Carboniferous.

In southwestern British Columbia, other Carboniferous series have been described, such as the Hozameen group and the Chilliwack group (map 737*A*, Canada Dept. Mines and Resources). The Hozameen is principally an intimate association of andesitic lavas (greenstone) and ribbon chert with argillaceous partings. No fossils have been found in it, but the rocks are similar to much of the widespread Cache Creek series, from which a number of collections of Permian fossils have been made in recent years. The Hozameen beds are probably somewhat younger than those of the Chilliwack group as exposed along the International Boundary. This series is predominantly sedimentary and has yielded marine Upper Carboniferous fossils.

Black slates, cherts, and limestones of Ordovician age occur in the northern Cascade Mountains,[52] as well as Carboniferous rocks equivalent to the Cache Creek group. Devonian rocks have been described west of the Cascades on the San Juan Islands.

Though major parts of the large area just described in northern Washington and southern British Columbia are underlain by highly metamorphosed crystalline rocks, they can no longer be regarded as entirely "Archean" or pre-Cambrian. Cairnes[53] and others have rather conclusively established their origin as metamorphic associations of the great Jurassic batholiths and their present state of metamorphism as no indication of the time when they were deposited. The formations mentioned above give a fragmentary conception of the nature of the accumulations in the Cordilleran geosyncline in this region in pre-Mesozoic time. Evidence and opinion seem to be growing to shape the conclusion that a large part of the Paleozoic rocks were deposited in Carboniferous time and that the Permian especially was a time of widespread volcanic accumulations.

A point of interest is the association, practically everywhere in the areas mentioned, of the Jurassic intrusions and the Carboniferous volcanics. The coincidence of the belt of intrusions and the earlier volcanic accumulations is shown on the map of Figure 6.

In southeastern Alaska.—The Paleozoic rocks in southeastern Alaska from latitude 54°30′ to 60° N. are of geosynclinal thickness and constitute a number of formations of Ordovician, Silurian, Devonian, Mississippian, Pennsylvanian, and Permian ages.[54] The stratigraphic succession is given in Table 1.

One of the commonest types of rock is andesite in various forms. It occurs in at least seven formations from Permian to Ordovician age, and perhaps older. Many of the volcanic rocks are now greenstone schist. Pillow lava is abundant in the Lower and Middle Ordovician, Silurian, Middle and Upper Devonian, Lower Permian, and Upper Triassic.

The other predominant rock types are sheared graywacke, slate, and phyllite. The vast amount of greenish graywacke with associated slate is the most striking feature of the stratigraphic sequence of southeastern Alaska. Graywacke is found in every system of the Paleozoic and Mesozoic, and in many places it is diffi-

[52] W. S. Smith, "Stratigraphy of the Skykomish Basin, Washington," *Jour. Geol.*, Vol. XXIV (1916), pp. 559–82.

[53] See ftn. 51.

[54] A. F. Buddington, and T. Chapin, "Geology and Mineral Deposits of Southeastern Alaska," *U.S. Geol. Surv. Bull. 800* (1929).

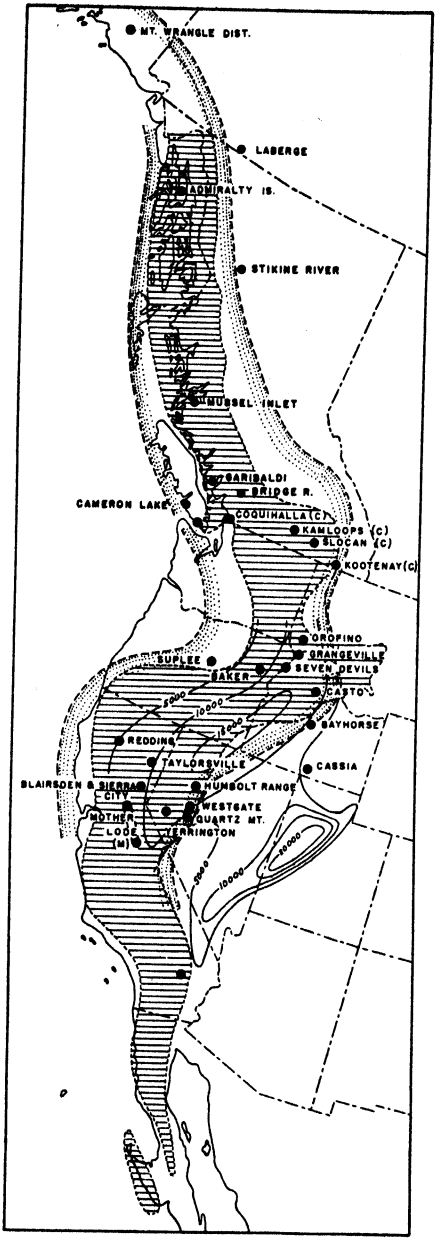


FIG. 6.—Map showing coincidence of Permian volcanic trough (*stippled margins*) and zone of Nevadan intrusives (*lined*). Dots indicate location of Carboniferous, Permian, and Triassic areas referred to in the text, Pennsylvanian and Permian basins combined isopached. Zone of Nevadan intrusives includes nearly all satellites and palingenetic areas.

TABLE 1

Series	Character	Thickness (Feet)
Jurassic		
	—Unconformity—	
Upper Triassic	Andesitic rocks, including breccia with limestone matrix and lava flows (in part with pillow structure), locally interbedded with slate and other sediments	1,400+
	—Unconformity—	
	Conglomerate, sandstone, and limestone; in the Ketchikan district includes considerable black slate in upper part	1,600±
	—Unconformity—	
Permian	Thick-bedded limestone: with common to abundant intercalated layers of white chert	1,000
	Conglomerate, limestone, sandstone, andesitic and basaltic lava, tuff, and, locally, rhyolitic volcanic rocks	3,000±
	—Unconformity—	
Pennsylvanian(?)	White massive limestone	100+
Mississippian	Interbedded coarsely crystalline limestone and black chert, overlain by interlayered dense gray quartzite and cherty limestone; sparse conglomerate	1,000
Upper Devonian	Basalt, andesite (in part pillow lava), tuff, limestone, sandstone, slate, and conglomerate	1,000
	—Unconformity(?)—	
Middle Devonian	Limestone	600+
	Andesitic green to gray tuff (locally cherty) and graywacke, with, locally, fine conglomeratic layers, intercalated limestone, and a minor amount of andesitic lava and breccia	2,400+
	Andesitic lava (in part pillow lava), breccia, tuff, conglomerate, and, locally, rhyolitic lava	2,000
	Interbedded limestone, slate, chert, andesitic lava, breccia, tuff, and, locally, conglomerate	?
	Conglomerate and graywacke-like sandstone, with, locally, interbedded limestone	2,000
	—Unconformity—	
Silurian	Green-gray graywacke with sparse conglomerate beds; interbedded red, green-gray, and gray graywacke-like sandstone, with a small amount of shale	5,000+
	Green-gray shale with intercalated red beds and thin-layered fine-grained gray sandstone, shale, and dense limestone	500+
	Predominantly thick-bedded dense limestone; intercalated with thick beds of coarse conglomerate, thin-layered limestone, nodular and shaly argillaceous limestone, and sandstone; limestone, 3,000± feet; conglomerate, 1,500 feet	4,500±
	Andesite (in part pillow lava) and andesite porphyry lava; conglomerate; with some associated graywacke, tuff, breccia, and limestone	3,000±
	—Unconformity(?)—	
	Indurated graywacke with associated black slate and sparse conglomerate and limy sediments	?
	—Unconformity(?)—	
Middle Ordovician	Indurated graywacke with associated black slate and sparse conglomerate and limy beds; locally, andesitic pillow lava and other volcanic rocks	?
Lower Ordovician	Thin-layered black chert with black graptolitic slate partings, graywacke, and, locally, andesitic volcanic rocks	?
Probably pre-Ordovician to Devonian Wales group (metamorphic rocks)	Greenstone schist with intercalated or interbedded limestone Limestone Schist with beds of limestone and slate Schist	?

After Buddington and Chapin (ftn. 54)

cult or impossible to tell one graywacke unit from another.

Limestone forms a very considerable part of each Paleozoic formation except the Ordovician. The thickest unit is in the Upper Silurian and is a very high-calcite variety. Some limestone carries considerable chert.

Beds of cobble and boulder conglomerate form conspicuous and thick members of the Silurian and Devonian formations. A peculiar, but common, form is composed of andesite and limestone pebbles and cobbles in a tuffaceous matrix. The same lithology is found in the Middle Devonian. Coarse conglomerate beds occur at the base of the Devonian.

Another characteristic lithology in the Paleozoic systems in southeastern Alaska is coarse, waterworn, intra-formational limestone conglomerate. Beds occur in the Silurian, Devonian, Permian, and Triassic formations; and in all of them the cobbles of limestone carry the same fauna as the formation in which the conglomerate occurs. Buddington believes the intra-formational conglomerates originated from crustal movements accompanying the volcanic activity during these periods.

Black slate and argillite are widely distributed, and thin-layered black chert, several hundred feet thick, occurs in the Ordovician and Mississippian formations. Thick-bedded chert and cherty tuff occur in the Middle Devonian, and white chert is common in the Upper Permian.

Schists and gneisses are also common and are the result principally of permeating hot solutions attendant upon the emplacement and solidification of the vast volume of magma, in addition to orogenic stresses.

In northern British Columbia and the Yukon.—The new geologic map of the Dominion of Canada (1946) summarizes what is known of the distribution of Paleozoic rocks in northern British Columbia and the Yukon. Great areas are still marked "Paleozoic, mainly sedimentary rocks," but other large areas are labeled "Carboniferous and Permian sedimentary and volcanic rocks." A report on the Cassiar Mountains, in the Finlay River district between latitudes 56 and 58 and longitudes 124 and 126, by Victor Dolmage,[55] describes a series of metamorphosed rocks of Carboniferous age. They are "green ash rocks pressed and altered into schists, interbedded with layers of graywacke, felsite, hälle-flinta, serpentine, and argillite." Along Takla and Stuart lakes and vicinity the series is made up of limestones, argillites, cherty quartzites, green schists, slates, volcanic flows, tuffs and breccias, and narrow bands of dolomite. *Fusulina* and other Carboniferous fossils have been found in some of these beds.

Underlying the Carboniferous series, great belts of schist and quartzite occur. Quartz mica schist constitutes about three-fourths of the whole. In many places the schist grades into quartzite, both of which were derived undoubtedly from siliceous sediments.[56] Such rocks as these are widespread and have been correlated with the Shuswap terrane of southern British Columbia, which now, as previously explained, is believed to be made up of rocks of several Paleozoic periods as well as pre-Cambrian. Also some coarse quartzites, quartz-pebble conglomerates, and limestones have been likened to the Cambrian strata of the southern Canadian Rockies, previously described.

The areas of such rocks are shown on

[55] "Finlay River District, B.C.," *Can. Geol. Surv. Sum. Rept.* (1927), Part A, pp. 19–41.

[56] *Ibid.*

the map of Figure 1. A great medial area of Proterozoic (Beltian?) rocks separates the western areas of Carboniferous rocks from the eastern Paleozoic rocks, but whether this was a highland in Paleozoic time or not is unknown. The relation of the Rocky Mountain trough to the Pacific, if at all distinct and separate, is not known in this far-northern Cordillera.

EVIDENCE OF PALEOZOIC OROGENY IN THE PACIFIC TROUGH

KLAMATH DIVISION

Extent.—In the Klamath division are included the late Paleozoic structures of the northern Sierra Nevada Mountains, the Klamath Mountains of northern California and southwestern Oregon, and the Ochoco and Blue mountains of central and eastern Oregon. These parts of the Klamath division were once connected but now are separated by extensive covers of later strata. The volcanics and most of the other clastics could have come only from the west, and it seems that the orogenic belt that supported some of the volcanoes and east of which submarine volcanic flows occurred must have lain a short distance west of the present exposures. The Klamath division was probably continuous with the Okanogan division on the north, from which it is now separated by the extensive Tertiary Columbia Plateau lavas.

Pre-Middle Devonian phase.—Evidence of crustal deformation in the Klamath system in upper Paleozoic time is the angular unconformity between the Copley meta-andesite and the Middle Devonian Kennett formation. The deformation was sufficient to fold the Copley and pre-Copley beds and to result in a landmass that was considerably eroded.

Since the overlying Kennett consisted originally of black clays for the most part, it is evident that the disturbance that resulted in the unconformity preceded the deposition of the fine clastics sufficiently to allow time for considerable erosion and for gentle submergence. Upon submergence of the Klamath area, the western landmass from which the sediments were derived must have been low lying or rather distantly removed, with a sea to the east, on the floor of which the black clays were deposited.

The disturbance is manifest in a gentle form several hundred miles to the east in the Nevada-Idaho Basin where, in the vicinity of the Manhattan geanticline (see Fig. 3, sec. *B–B′*), the Devonian rests on Ordovician. The Devonian Basin sank to its deepest extent in the central Nevada area and received chiefly lime sediments, as would be expected with a western landmass and an intermediate zone of accumulation of clastics.

The Copley andesites were slightly metamorphosed before the Jurassic intrusions superimposed local alterations; but whether the primary metamorphism was a result of the pre–Middle Devonian disturbance or others that followed in Carboniferous and Triassic time is not yet known. Hinds[57] points out that the degree of regional metamorphism on the Abrams and Salmon schists is considerably greater than that on the overlying strongly folded Chanchelulla and that an angular unconformity exists between them. If this relation is regionally true, it cannot be denied that a fairly intense orogenic disturbance separated the two formations. Regarding the post-Chanchelulla formations, however, it appears evident that geologists working in the Klamaths have not distinguished any special difference in degree of the original

[57] See ftn. 26.

dynamic metamorphism. The Copley meta-andesites, and the Kennett slates have about the same amount of metamorphism.

Late Devonian–early Mississippian phase.—Since the Bragdon formation of Mississippian age rests on both the Kennett and the Copley formations, it appears that the crust was restless again in late Devonian or early Mississippian time. The Bragdon was originally mostly clay, like the underlying Kennett, and it follows that the source of the sediments was a landmass, either low lying or distant, and that disturbance and erosion preceded the deposition of the Bragdon clays by some time.

The Bass Mountain basalt is the third episode of volcanism recorded to date in the Klamath system. The extrusion of lavas and tuffs on a large scale followed the pre-Mississippian disturbance and probably indicates continued crustal movements to the west.

Pennsylvanian phase.—In the Klamath Mountains the Lower Permian strata rest on the Mississippian if the latest age determinations are correct, and therefore the area was above water and suffered erosion during Pennsylvanian time. The uplift seemed more vigorous, yet the landmass was shorter lived, in the nearby Ochoco Mountains because Pennsylvanian beds there were deposited unconformably on older beds. Near Baker in eastern Oregon the Pennsylvanian argillites probably rest on the Burnt Ridge schist, which seems to be considerably older.

During most of the time that the Elkhorn Ridge (Pennsylvanian) was being deposited, much volcanic activity was occurring on the west, and pyroclastic contributions to the subsiding Pennsylvanian basin were intermittent but abundant.

In the Suplee district of central Oregon a slight angular unconformity separates the Pennsylvanian strata from the lower Permian, and therefore in places it seems that movements continued to the close of the Pennsylvanian.

Permian phase.—The Permian volcanics, which are so widespread in the great Cordilleran geosyncline, indicate that a fringe of volcanoes lay to the west, with submarine flows in the belt of present exposures and, unquestionably, that orogenic movements created either sizable landmasses or island archipelagos. The extensive Permian volcanism, together with the inferred crustal movements, is the first part of the Permian phase. Volcanism continued energetically in places into Triassic time, but in the Suplee district of Oregon an angular unconformity has been recognized between the Permian and the Triassic volcanic series. Compression in pre-Triassic time resulted in a series of folds striking northeast and north-northeast. Dips of 60° and more indicate that the folds were close and the deformation intense.[58] The cross sections (Fig. 5) show these folds in the Carboniferous and Permian strata, as well as the unconformity between them and the Triassic volcanics. The syncline of the Triassic beds strikes about N. 30° W. at the south end of its exposure and veers more northerly at its north end. The dips of the Triassic beds are greater than 40° in a number of exposures, but generally the folds seem more open than those of the beds below the unconformity. The episode of compression that folded the Triassic beds may be related to the late Jurassic Nevadan orogeny, or it may predate this event. Evidence is not at hand to make accurate dating possible.

[58] See ftn. 39.

In the Baker area, folding has been intense in all three formations exposed. The trends are westerly, the dips steep, folding probably isoclinal, and metamorphism dominant. Unfortunately, the critical lines of contact are obliterated by intrusions or masked by overlying Tertiary rocks, and, as the Burnt River schist is of uncertain age, it is impossible to work out satisfactorily the structural relations.[59] On the supposition that the Burnt River schist is pre-Carboniferous, the structure may be that of the north limb of a large anticlinorium, or possibly the schist may be overthrust on the Elkhorn Ridge argillite. It is also possible, in view of the notable shearing that all the rocks have undergone, that the Elkhorn Ridge argillite is thrust northward on the Clover Creek greenstone.

All the pre-Tertiary rocks except the youngest and most voluminous type of the intrusive bodies—the biotite-quartz diorite—have been subjected to thorough dynamic metamorphism. All have undergone shearing of the most pervasive sort and, concomitantly, have suffered mineralogic transformations. The metamorphism induced by the igneous intrusions has been slight. The Permian greenstones have been altered, where invaded, to hornfels or garnet gneiss, the Elkhorn Ridge argillite to spotted andalusite hornfels, and the Burnt River schist to amphibolite.[60]

The age of the dynamic metamorphism is the subject of greatest interest at this point. Is it post-Permian and pre-Triassic, late Triassic, or early Jurassic?

A group of basic intrusions in the Baker quadrangle are all believed to be magmatically related and hence of a single intrusive cycle. They are all affected equally by the dynamic metamorphism. Intrusive rocks, mentioned as basic plutonics, were intruded into Upper Triassic sedimentaries in the Ochoco Range and were exposed by erosion before being covered by the Lower Jurassic strata.[61] The basic plutonics have been altered to serpentine. If the serpentinization that Lupher mentions is of dynamic origin, as it might be,[62] then it would appear to be related to the metamorphism in the Baker quadrangle, and the general metamorphism of the prediorite rocks would be of late Triassic age. At least it seems safe to conclude that the folding, thrusting(?), and metamorphism is of post–early Permian and pre–early Jurassic time.

Writing about the Calaveras in the Mother Lode area, Knopf[63] states that the metamorphism that affected the formation differs markedly from that which altered the associated, but much younger, Upper Jurassic Mariposa rocks. He concludes that the metamorphism of the Calaveras took place at an earlier time and under other conditions than those that prevailed during the crustal folding in late Jurassic time. He believes that this pre-Mariposa metamorphism was a result of dynamic crustal compression. At the time that the Calaveras was being metamorphosed, or shortly afterward, considerable masses of diorite were intruded. The diorite is now, for the most part, schistose or gneissic, as a result of cataclastic deformation.

Spatial and time aspects of the orogenic belt.—The evidence summarized on the past several pages regarding the Paleo-

[59] See ftn. 41.

[60] *Ibid.*

[61] R. L. Lupher, "Jurassic Stratigraphy of Central Oregon," *Proc. Geol. Soc. Amer.*, Vol. LII (1941), pp. 219–70.

[62] L. V. Pirsson and Adolph Knopf, *Rocks and Rock Minerals* (2d ed.; New York: John Wiley & Sons, 1926).

[63] See ftn. 37.

zoic Cordilleran geosyncline and the Klamath division points to the following conclusions. A western belt of volcanoes contributed prominently to the accumulations in a flanking trough on the east. At certain times of little or no volcanic activity the land areas were low or distant. Crustal unrest was manifest from time to time in the volcanic belt, and the zone of disturbance extended eastward into the geosynclinal area sufficiently to cause unconformities between the sedimentary series. Not until late Permian or Triassic time, however, did intense orogeny pervade the Cordilleran geosyncline and cause the folding and metamorphism that are found in the Klamath division.

The different trends of the pre-Triassic and post-Triassic folds in the Suplee district and the still different strike of the beds in the Baker district may be due to orogenies of different times whose structures were discordantly superposed on one another. Also, as will be seen later, the Oregon part of the Klamath division and the Okanogan division, next to be discussed, are at the intersection of two great arcs of the general Paleozoic volcanic and orogenic belt, and in this area of intersection the structural trends would undoubtedly be complex. It is certainly the area in which the Nevadan batholiths are most irregular and complex and extend farthest eastward (see maps, Figs. 1 and 6). The palingenetic processes also were by far the most extensive at this area of intersection.

OKANOGAN DIVISION

The Okanogan division of the Pacific trough designates the Paleozoic structures in Washington and southern British Columbia. The areas of outcrop have already been listed. According to Krauskopf,[64] the phyllites, greenstones, and chlorite schists of the Okanogan Valley in Washington are regionally metamorphosed, and on this earlier metamorphism is superposed the high-grade contact metamorphism of the intrusions.

In British Columbia to the north the metamorphism of the Shuswap rocks is considered Jurassic and early Cretaceous in age and was caused by the intrusions and associated orogeny, but it is likely that one or more episodes of dynamic metamorphism preceded the contact metamorphism. In correspondence Cairnes says:

> There is little doubt that such Carboniferous formations as the Cache Creek, Anarchist, Chilliwack, Hozameen, and Kobau were folded and subjected to some metamorphism prior to the main period of batholith intrusion. The intrusive bodies definitely cross-cut the folded structures and are themselves relatively fresh and unstrained. However, it is quite uncertain as to whether the Carboniferous strata were subjected to one or more deformations. It seems reasonably certain that, if more than one, the last occurred in Cretaceous or in very late Jurassic time, as it has affected intrusive bodies, such as the Custer granite-gneiss and the Osoyoos and Fairview batholiths of late Jurassic age, presumably. The Osoyoos and Fairview batholiths are involved in the folding and metamorphism of the Kobau group and are definitely older than the bulk of the batholiths of the area.
>
> In southwest British Columbia it seems generally true that formations of Triassic and earlier age are more severely deformed and more completely altered that those of Jurassic or later Mesozoic times, and perhaps the consensus of opinion would be that this early deformation was the result of an orogeny forerunning the emplacement of the enormous quantities of batholithic materials in the late Jurassic and succeeding Cretaceous and Tertiary periods. Where fossiliferous evidence is lacking, a principal difficulty in this part of the province has been to separate, either lithologically or structurally, the Triassic from late Paleozoic formations, rather than the Triassic from those of later age. Farther north, in central British Columbia, the situation seems different. There Permian and Carboniferous

[64] See ftn. 46.

formations are more deformed than the Triassic (probably all Upper), Jurassic, and early Cretaceous rocks. The difference is probably related quite closely to a variable history of batholithic intrusion. In the south, the principal periods of intrusion seem to have been in later Jurassic, post-Lower Cretaceous, and mid-Tertiary times; in the north there are reasons for believing that the earliest principal period of intrusion was in Triassic (probably early or pre-Upper) time, and others of consequence in pre-Upper Cretaceous and pre-Eocene (Laramide) times. Much of the information however, is speculative to say the least, and there is an aggravating lack of sedimentation where stratigraphy could be of most help, a dearth of key fossils where they would be most useful, and an inconsistency in the interpretation of plant and invertebrate remains from the same beds that is little short of exasperating.

B. B. Brock[65] and James Gilluly[66] both believe the metamorphism of the Shuswap rocks to be of dynamic origin. Some petrologists believe that the presence of magmatic solutions is necessary to produce a schist, and, if so, the dynamic metamorphism of Gilluly may be related to the Jurassic intrusions. The problem of pre-Jurassic orogeny, therefore, is not clear as a result of the study. Cairnes[67] concludes that the foliation of the Shuswap is primarily an inherited structure, conformable with the original bedding, and that prevailing structural trends, as indicated by the foliation, may some day serve a useful purpose in demarcating separate formations whose age and structural history may differ. He points out that the structures on the whole are gentle and that this is a diagnostic feature of the terrane and bears directly on the geologic history, namely, that they were not much folded before or during the granitic intrusions.

[65] "The Metamorphism of the Shuswap Terrane of British Columbia," *Jour. Geol.*, Vol. XLII (1934), pp. 673–99.

[66] "Mineral Orientation in Some Rocks of the Shuswap Terrane as a Clue to Their Metamorphism," *Amer. Jour. Sci.*, Vol. XXVIII (5th ser., 1934), pp. 182–201.

[67] See ftn. 51.

ALEXANDER DIVISION

Extent.—The Alexander division of the Pacific trough is named for the Alexander Archipelago of southeastern Alaska, but the division includes, besides the islands, parts of the crust along the mainland, now largely occupied by the great Coast Range batholith. Paleozoic rocks are exposed from the St. Elias Range southward to Dixon Entrance; but from there to the south end of Vancouver Island only Mesozoic rocks crop out. The batholith separates the sediments of the Alexander system from Paleozoic and Mesozoic rocks to the east, which are as yet little studied but probably are, in part, of the Pacific trough facies and, in part, of the Rocky Mountain trough facies.

Pre-Middle Devonian phase.—The relations of the pre-Middle Devonian formations to one another are everywhere doubtful.[68] The Wales group, part of which seems to be of pre-Ordovician age, is overlain in places by beds of Lower and Upper Silurian age, but whether conformably or unconformably is not known. The Wales rocks are relatively much more deformed and metamorphosed than the known Silurian formations, but it is possible that this difference may be due to excessive metamorphism resulting from the intrusion of large masses of igneous rock that are only slightly exposed at the surface. Wales rocks have not been observed in contact with rocks of known Ordovician age, and the relations between them are not known. The Wales rocks were originally sandstone, shale, lava, and tuff.

[68] See ftn. 54.

Graywacke and slate formations of the Ordovician and Lower Silurian are isoclinally folded with nearly vertical dips wherever found, regardless of the effects of the igneous intrusions. The Upper Silurian beds, on the other hand, are metamorphosed and folded locally in connection with the igneous masses but, in general, not so strongly as the underlying graywacke and slate. There are two possible explanations for this: the Ordovician and Lower Silurian beds may have been folded before the deposition of the Upper Silurian, or the more intense folding of the older beds has resulted from folding of the incompetent graywacke and slate beneath arches of the thick, massive Silurian limestone, which acted as a competent bed. There is no doubt that the difference in competency has been an important factor in producing the difference in folding between the two groups of rocks, but whether it is the only factor is questionable.[69]

The volcanics, the graywackes, the conglomerates, the suggested unconformities, and perhaps the difference in degree of folding and metamorphism of some of the formations all point to the existence and close vicinity of an orogenic belt, where movements were occurring. The belt of major orogeny lay to the west, and the site of preponderant deposition was in the trough, to which, recurrently perhaps, the crustal movements spread and disturbed the sediments and lava flows.

Middle Devonian phase.—Middle Devonian formations overlie the Lower and Upper Silurian beds unconformably. The relation is clearly exposed on several islands in the archipelago.[70] The Middle Devonian strata in the Klamaths, it will be recalled, overlie the older formations unconformably. The identification of a crustal disturbance at the same time at both places, even though its details remain unknown, emphasizes the magnitude of the orogeny represented by the unconformity.

Some suggestion of an unconformity between the Middle and the Upper Devonian is noted at two places (Coronados Islands—a small group at the mouth of Port St. Nicholas–Prince of Wales Island—and small islands in Trocadero Bay).

Permian phase.—In the Ketchikan district the limestone division of the Permian overlies the Upper Devonian unconformably. In the Keku Islands of the Petersburg district the lower division of the Permian seems to be unconformable on formations as old as Silurian. At the east end of Suemez Island the Permian appears to rest unconformably on the Upper Devonian.[71]

Lower Triassic phase.—The basal conglomerates of the Upper Triassic formations in the Alexander Archipelago at many localities carry abundant fossiliferous limestone cobbles of Devonian and Permian age. At most places the Upper Triassic beds overlie the upper division of the Permian; but locally they overlie Devonian formations.[72]

A basal Triassic conglomerate has been observed in a number of districts in British Columbia and the Yukon, in addition to southeastern Alaska. S. J. Schofield[73] has listed the places as follows: Cameron Lake on Vancouver Island, Mussel Inlet, Garibaldi region, Britannia map area, Coquehalla, Bridge River, Stikine River, Laberge area, Slo-

[69] *Ibid.*

[70] *Ibid.*

[71] *Ibid.*

[72] *Ibid.*

[73] "Cascadia," *Amer. Jour. Sci.*, Vol. CCXXXIX (1941), pp. 701–14.

can, Kootenay, and Kamloops (see map, Fig. 6).

The conglomerate consists of metavolcanics, mica schist, metaquartzite, diorite or gabbro, and, very persistently, granite, granodiorite, and quartz-diorite pebbles and boulders. In several places it is massive, with boulders up to 8 inches in diameter. In the Laberge area the massive conglomerate is 700 feet thick and consists of well-rounded boulders, 2 inches–2 feet in diameter, and predominantly of volcanic porphyry and graywacke. Granodiorite boulders are abundant, and limestone boulders are scattered. The silicic plutonic boulders could not have been derived from the Jurassic batholith because the Triassic conglomerates are cut and nearly everywhere metamorphosed by them.

Arkose occurs with the conglomerate in the Britannia map area. The lack of rounding of the grains, the freshness of the plagioclase, and the thickness of the unstratified beds have been taken to signify rapid accumulation and a near-by source. The conglomerates probably indicate the same conditions. Of special interest is the observation that the near-by source was a land in which silicic plutonic rocks were exposed.

The Wrights[74] postulated a northeasterly trend to the folds, which they believed was the dominant structure at the close of the Paleozoic over all southeastern Alaska. The present structural trends are northwesterly, and, according to Buddington,[75] the Wright concept would need, in the face of its exceptional nature, special proof. Buddington saw evidence of northeasterly folds at only one place, and he does not believe that the contention is proved. Nevertheless, he believes that the unconformable relations of the Upper Triassic sediments and the older ones indicate a time of marked disturbance, perhaps of major orogeny. Schofield points out that the Triassic beds in places have the same strike and dip as the underlying ones and that, although in those places the early Triassic orogeny seems to lose importance, the theory of divergence in trend of folds of an earlier orogeny with those of post-Triassic orogeny is not supported. The basal Triassic conglomerates reflect orogeny in a near-by area, and the absence in places of an angular unconformity where the conglomerates were deposited does not minimize the importance of the orogeny.

VOLCANIC ARCHIPELAGO

CONCEPT OF A WESTERN LANDMASS

In the opinion of several geologists, the source of the sediments, and especially of the volcanics, of the Cordilleran geosyncline in Paleozoic time was to the west. Schuchert named the landmass from which they came "Cascadis" (*Historical Geology* [1924]) and characterized it as the greatest of all "borderlands." In the Mesozoic era he believed Cascadis was divided into two smaller landmasses, each with shorelines farther west than the Paleozoic one. The Schuchert concept of Cascadis seems to be the most widely accepted theory today among paleontologists and stratigraphers. Deiss[76] recently depicted the shoreline of Cascadis during the Cambrian in about the position that Schuchert illustrates it, namely, at the western margin of the southern Rocky Mountain trough, the Montana dome, and the northern Rocky Mountain trough (see map, Fig. 1).

[74] F. E. and C. W. Wright, "Lode Mining in Southeastern Alaska," *U.S. Geol. Surv. Bull. 284* (1906), pp. 30–54.

[75] See ftn. 54.

[76] See ftn. 4.

Schofield[77] has reviewed the concept of "Cascadia" and agrees that a western landmass existed; but, on the basis of Proterozoic, Silurian, and Triassic conglomerates of British Columbia and southeastern Alaska, he regards it as having lain west of the present west coast of North America, except during the Proterozoic (Beltian), when, he thinks, the shoreline was just west of Kootenay. Cairnes[78] has shown that the Shuswap terrane is not all pre-Cambrian rock and therefore not a former land area, but that, on the contrary, it is the metamorphosed facies of rocks of several periods and that it was the site of deposition during at least part of the Paleozoic. The Shuswap terrane has been depicted as part of Cascadis by Schuchert and others, but this seems now disproved.

The previous pages on the great Cordilleran geosyncline—its divisions, sediments, and deformation—have suggested several stimulating lines of evidence by which the position and constitution of the lands from which the sediments came can be deciphered. The nature of the sediments alone is very informative; and this, together with their distribution, provides enough data to reconstruct a general picture of the western lands.

EVIDENCE OF A WEST-LYING ARCHIPELAGO

Volcanic rocks.—The abundance of volcanic rocks, mostly tuffs and lavas, in the Pacific trough, with none in the eastern, has been noted. The volcanoes were not situated along a landmass between the two troughs because such an elevation probably did not exist except in the south in Carboniferous time, when the Manhattan geanticline rose. If a medial land area had existed and had been the site of volcanoes, the eastern trough would have received as much volcanic material as the western. It is therefore concluded that the volcanoes were aligned along the western flank of the Pacific trough. They are known to have been especially numerous opposite the Klamath and Alexander troughs and were active at intervals in each period from pre-Ordovician to the Present.

In the modern geologic picture, rows of active volcanoes are on the concave side of island arcs or archipelagos, but on the larger islands, such as Japan, the volcanic rocks are widely distributed on both sides. It therefore seems of some significance that the Klamath and Alexander troughs are on the concave sides of what would be the west-lying island arcs.

The great bulk of the lavas in both the Klamath and the Alexander troughs are andesites. Some basalts occur in the Klamaths, but only andesitic lavas have been found in the Alexander trough. This is especially instructive because the volcanoes of the island arcs, both modern and ancient, have emitted chiefly andesites. W. H. Hobbs[79] has recently reviewed this distribution and has drawn an andesite-basalt line which separates the basalts of the volcanoes of the ocean floor from the andesites of the island arcs in the western and southern Pacific. Rarely do appreciable quantities of andesite occur on the mainland opposite an archipelago. Inspection of the geologic map of eastern Asia (Fig. 7) reveals abundant andesite on the Japanese archipelago but an absence of it on the Asiatic mainland across the Sea of Japan and the Sea of Okhotsk.

It may be concluded, therefore, that the andesitic nature of the preponderance of volcanic material in the western trough of the Cordilleran geosyncline

[77] See ftn. 73.

[78] See ftn. 51.

[79] "Mountain Growth: A Study of the Southwestern Pacific Region," *Proc. Amer. Phil. Soc.*, Vol. LXXXVIII (1944), pp. 221–68.

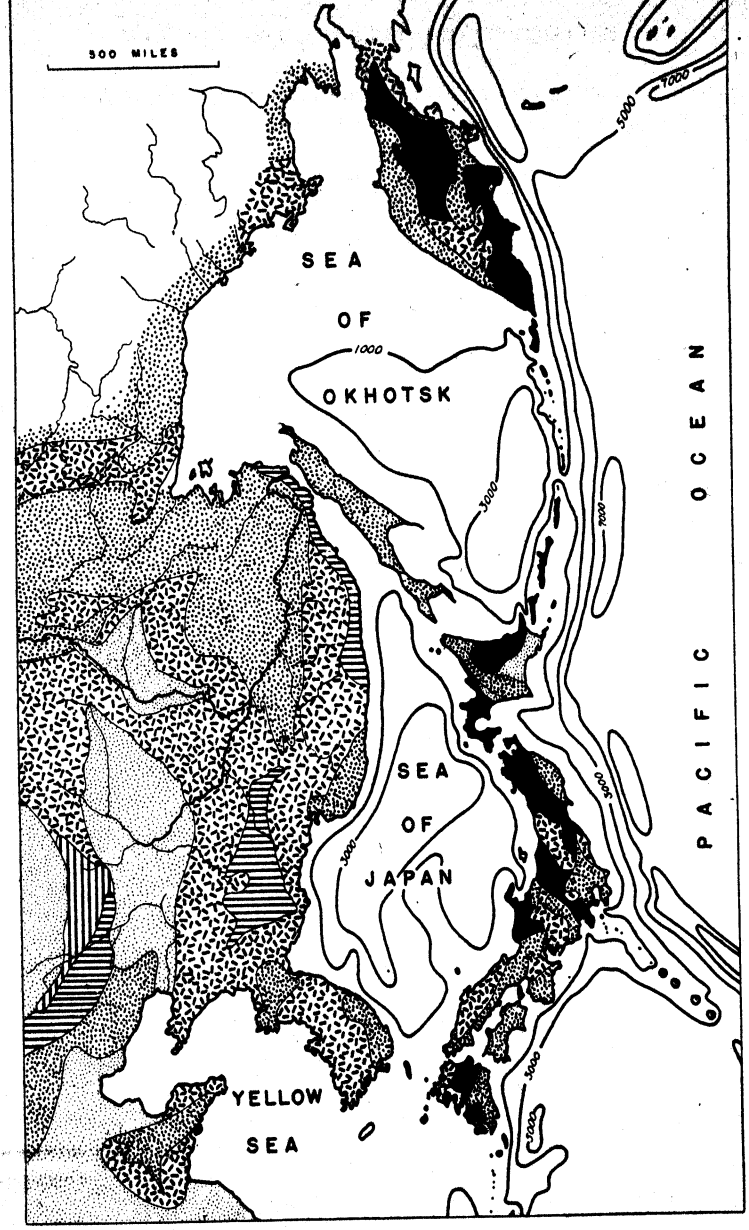


FIG. 7.—Generalized geologic map of the Japanese Archipelago and the eastern part of Asia. Isobaths are in meters. Coarsely stippled areas are those chiefly of sedimentary rocks, but with large areas of Archean gneisses and schists and some smaller areas of intrusive and extrusive rock. Finely stippled areas denote alluvium. Hachured areas as those of plutonic rocks, chiefly granite to granodiorite, but with considerable areas of Archean gneiss and schist and some sedimentary rocks. Solid black areas are andesite. Horizontally ruled is basalt and vertically ruled is trachyte.

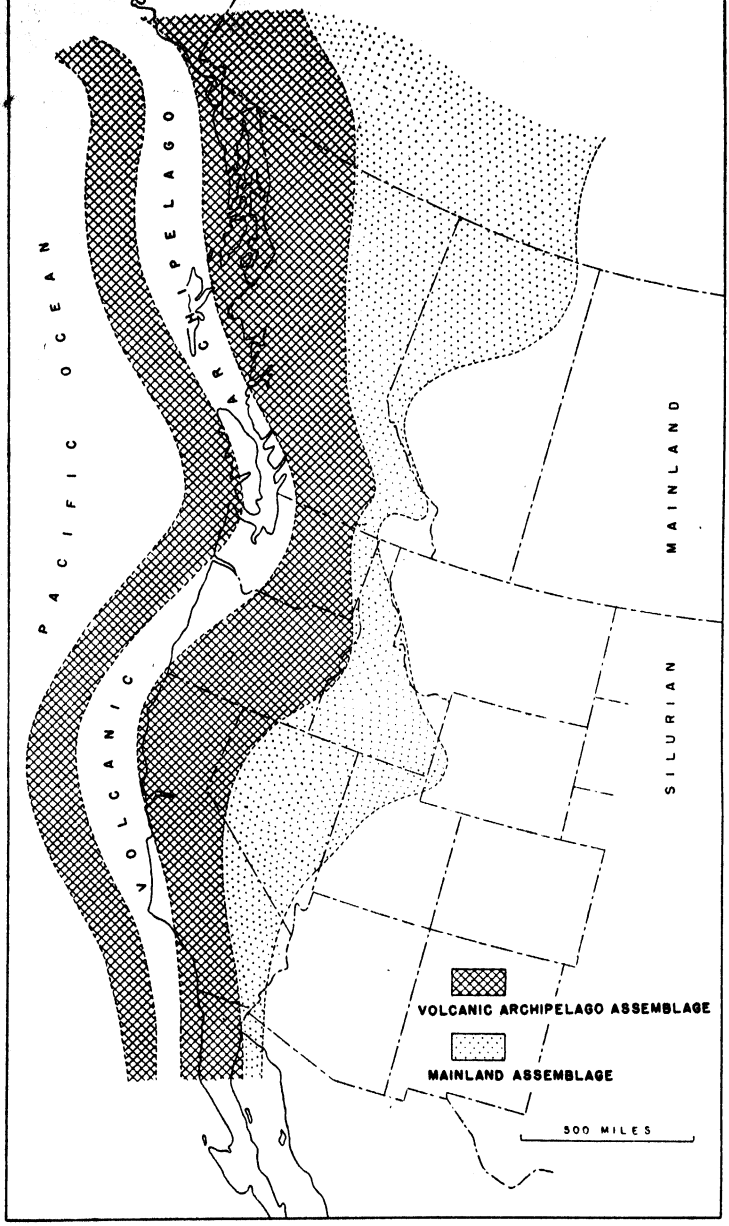


FIG. 8.—The volcanic archipelago and its relation to the sediments of the Paleozoic Cordilleran geosyncline. The map is generalized for the Silurian period on the assumption that the Copley volcanics of the Klamath Mountains are Silurian in age and correlate with the Silurian volcanics and graywackes of the Alexander Archipelago. The volcanic archipelago assemblage consists of andesitic volcanics in many forms, graywacke, conglomerate of andesite and granitoid rocks, massive limestones and coarse intra-formational limestone conglomerates, chert, arkose, dark shale, and siliceous shale (all the above varieties are generally found now in their metamorphosed form). The mainland assemblage consists of orthoquartzite, arkosic sandstone, sandstone, conglomerate (generally pea and pebble size, composed of quartzite and quartz pebbles), limestone and dolomite, cherty limestone and chert, and various shales and siltstones.

also indicates that the land to the west was a volcanic archipelago.

Graywackes, arkoses, and conglomerates.—A graywacke is a dirty sandstone in the sense that a clean sandstone is composed largely of quartz grains. Graywackes have an abundance of slate, phyllite, argillite, and chert fragments. The grains come from previously deposited clastics that have suffered low-grade metamorphism.[80] From a study of graywackes in the Appalachian geosyncline, from the Alpine molasse, and from the Siwalik strata of India, Krynine concludes that graywackes are due to erosion of previously deposited sediments in a geosyncline that had been folded and somewhat metamorphosed by widespread orogeny along the margins. He also thinks it is possible that geanticlines with geosynclines could bring the previously deposited sediments to the surface and result in the formation of graywackes.

Buddington[81] states that the Silurian graywackes, in general, of southeastern Alaska are composed of particles of rock similar to the kinds that form the pebbles and cobbles in the conglomerates with which they are interbedded and, in addition, of a considerable percentage of plagioclase, potassic feldspar, and quartz grains. The conglomerates are largely made of andesite pebbles and boulders; but slate, diorite, rhyolite, and limestone pebbles are abundant, if not dominant, in some conglomerates. One specimen of graywacke of Devonian or Silurian age, for example, consisted of particles of andesite, felsite, plagioclase, granophyre, quartz, spherulitic rhyolite, and orthoclase, with a chloritic and slightly calcareous groundmass.

[80] P. D. Krynine, "Paleogeographic and Tectonic Significance of Graywackes" (abstr.), *Bull. Geol. Soc. Amer.*, Vol. LII (1941), p. 1916.

[81] See ftn. 54.

The association of the graywackes and conglomerates that Buddington describes is very revealing of their origin and substantiates Krynine's conclusions. The conglomerates in themselves are indicative of a volcanic archipelago to the west and deserve further mention. The following is a résumé of the Silurian conglomerates according to Buddington. Varieties of conglomerate are as follows:

1. A conglomerate composed almost wholly of well-rounded andesite or andesite porphyry cobbles and boulders; the matrix may be calcareous, and lenses of limestone are intercalated, but limestone cobbles are sparse.

2. A conglomerate composed almost wholly of limestone cobbles or boulders in a limestone or andesitic tufflike matrix; this type is rare, but beds 100 feet thick have been noted.

3. Peculiar conglomerates intermediate between 1 and 2, consisting of pebbles and cobbles of andesite and limestone in a greenish tufflike matrix.

4. A homogeneous-appearing rock composed of fragments of andesite in a matrix of the same material; the structure is that of a conglomerate or waterworn breccia.

The limestone fragments are usually of a dense-textured limestone typical of the Silurian, and many carry fossils of Silurian age. The fossils are the same as from the overlying limestone. It is, therefore, believed that the limestone conglomerates are intra-formational and that the limestone fragments are of practically the same age as the volcanic fragments. Vertical movements of the sea bottom, perhaps local, must have accompanied the volcanism and must have resulted in contemporaneous erosion and submarine slumping of slightly compacted fine lime mud. A part of the volcanic material, at least, must have been

erupted from central volcanoes, which were built up above the surface of the ocean and were thus subjected to erosion.

Although recognizing unsolved elements in the problem of origin of the graywackes, conglomerates, and limy argillaceous beds, Buddington visualizes a sedimentary environment as follows; the great lens-shaped beds of conglomerate may be local deposits made by torrential streams, and the graywacke may be in part the more finely comminuted peripheral marine equivalent. The calcareous shale and argillaceous limy beds, which are locally intercalated with the clean, thick-bedded limestone, may be in part the more distant offshore equivalent of the conglomerate and graywacke.

The limestone is in part dense white on fresh surfaces and massive, with only rare, if any, evidence of stratification. Beds as thick as 2,000 feet have been observed. In part it is interbedded with thin-layered limestone, nodular and shaly limestone, calcareous shaly argillite, dense platy siliceous layers, green-gray shale, and sparse buff-weathering sandstone. The massive limestone seems to be due to rapid deposition, and, where clean, the site of accumulation was sufficiently distant from land so as not to have received any clastic material. Volcanic activity has been thought of as contributing to the deposition of the limestone through the activity of magmatic waters or meteoric waters draining from a volcanic terrane, but the chemistry and oceanography of the problem have not been worked out.

Schofield[82] has already been cited as having discussed the problem of granitoid pebbles and cobbles in the conglomerates of several periods, especially the Triassic. Buddington refers to them also. In one locality the Britannia map area of British Columbia, an arkose is described as composed of irregular grains of quartz, plagioclase, orthoclase, and sericite schist. The lack of rounding of the grains, the freshness of the plagioclase, and the considerable thickness of the unstratified beds prove that the material accumulated rapidly and was transported only a short distance from a source of granitoid plutonic rocks. Buddington failed to trace the granitoid clastics to their source, despite the fact that their size and abundance indicated to him a nearby local origin. It seems necessary, he believes, to assume that granitoid intrusions existed in a land that formerly stood to the west, where only the Pacific Ocean now lies.

Krynine[83] has studied the tectonic significance of arkoses and concludes that they are deposited just after a granitoid terrane has been uplifted and while it is being vigorously dissected. They are derived from the deformed geosyncline into which granitoid rocks have been intruded. The plutons become exposed by erosion of the mountains created by orogeny and then are uplifted in a further stage of deformation and vigorously eroded.

Granite plutons are seldom exposed in arcs of small volcanic islands. We must look to the larger islands of an archipelago for the source of granitoid conglomerates and arkoses. The geologic map of the Japanese Archipelago (Fig. 7) shows extensive areas of granitic intrusions and pre-Cambrian gneisses, which could furnish the necessary material. A major archipelago like the Japanese has had a long orogenic history and is composed of rocks that will make not only graywackes but also arkoses. Such a one seems to have been the source land of

[82] See ftn. 73.

[83] See ftn. 80.

the sediments of the western part of the Cordilleran geosyncline.

Orthoquartzite and chert.—The eastern trough lacks the volcanics, graywackes, and arkoses, but it has a dominant lithology that is exceedingly scarce in the western trough—sandstones cemented by silica. These are commonly called "quartzites," such as the Tintic, Brigham, Bingham, Weber, Wells, Quadrant, Flathead, Deadwood, etc.; but perhaps the names "orthoquartzite" and "quartzose sandstone" would safeguard against the misconception that they are metamorphic rocks. Some of the Proterozoic quartzites in the Rocky Mountain trough are regionally metamorphosed, but it is very doubtful if any of the Paleozoic and Mesozoic quartzites are affected in the least.

According to Krynine,[84] sedimentary orthoquartzites and quartzose sandstones originate either as deposits following prolonged chemical weathering of a terrane or though re-working of pre-existing quartzose sediments. The first group generally develops after chemical decay in peneplained regions. The shallow seas into which the sands were deposited were, at the time, precipitating silica, and, as a result, orthoquartzites were formed. If part of the sea bottom is removed from sand deposition, considerable beds of chert will form. If Krynine's postulate is true, it would appear that the sand grains of the orthoquartzites were derived from the mainland of the continent. It does not seem possible that they could have come from the west, where volcanic activity dominated. Associated with the Paleozoic quartzites are several limestone formations, 500–2,000 feet thick, and several somewhat thinner shale formations. The assemblage of quartzite, shale, and limestone supports the conclusion of a mainland origin.

Great beds of chert are present in the sediments of the volcanic archipelago. Extensive beds of chert and cherty limestone are present in the Rocky Mountain trough, as well as in the inland basins and shelfs of the mainland, and therefore the factors governing the precipitation of the silica are probably several. Its transportation in solution in marine currents may result in precipitation at a great distance from its source. I find it easy to believe that a large part of the silica originated in the volcanic activity of the archipelago; that some of it was carried by currents across the seas between the archipelago and the mainland free from the area of deposition of volcanic material; and that it was precipitated copiously in the shallow seas of the eastern trough and mainland shelf, where, from place to place and from time to time, clay, lime mud, or sand was accumulating. The problem needs special attention beyond the limits possible here, but the distribution of the cherts and orthoquartzites fits in nicely, if only theoretically, with the postulate of a western volcanic archipelago.

Distribution of troughs and sediments.—The pre-Devonian troughs of the Cordilleran geosyncline are isopached in Figure 1. The spread of data and the amount of interpretation have been discussed on previous pages. In brief review it is clear that the contouring of the western trough between the Alexander Archipelago and the Klamath Mountains is principally a guess, but, at least, a logical one and not at variance with known information. The Klamath trough is contoured on the assumption of a Paleozoic age of the Copley, Chanchelulla, and Salmon formations. The western flank of the northern Rocky Mountain trough is

[84] *Ibid.*

masked by metamorphism, so that its exact configuration is not known. Also the continuation of the trough northward from Jasper is fairly speculative. Elsewhere in the great geosyncline, substantial data support the contouring.

The pattern of the troughs, together with a postulated western volcanic archipelago, reminds one of the seas and archipelagos of the western Pacific, and it is natural to compare the two. In doing so it must be borne in mind that in our own western geosyncline we have the sedimentary record through the geologic periods, whereas in the western Pacific Ocean we have the shorelines of the moment, the distribution of the rocks on the archipelagos and the mainland adjacent to the seas, but practically no information about the sediments that are forming in the seas. A few dredgings have been made in a few places and have been reported on, but these represent at best only the topmost layer. Under the great seas between the archipelagos and the mainland we have little idea where troughs of greatest accumulation are or what the nature is of the sediments in the troughs below the surface. Instead of making a comparison, we find ourselves supplying the lost parts of one with the known parts of the other and making a complete geologic province, for we have the sedimentary record in one and the source of the sediments and the site of their deposition in the other. Knowing what we do of the origin of sediments, it is not difficult to recognize the fit or misfit of the parts.

In the first place, the shape and scale of the geosynclinal elements of the two regions is comparable (cf. Figs. 7 and 8). The Japanese Archipelago including Kamchatka was selected for comparison because it is the closest modern parallel that could be found. The curvature of the arcs, their convexity toward the Pacific, their width and length, are conspicuously alike.

In the second place, the geology of the Japanese Archipelago is somewhat the same as that postulated for the source land of the sediments of the Pacific trough of the Paleozoic Cordilleran geosyncline. If the Tertiary and Cretaceous strata are imagined removed from the Japanese islands, then the proper rocks and terrane are present to supply such sediments as are found in the Pacific trough. The most abundant rocks mapped in the Japanese Archipelago are as follows: andesite, granite, syenite, schistose granite, gneiss, schist, slate, chert, sandstone, limestone, diorite, pyroxenite, amphibolite, gabbro, and trachyte, in approximately decending order of abundance. From the nature and distribution of these rocks it is realized not only that the Japanese Archipelago and related seas are a comparable example but that their history probably ran through the Paleozoic era in a way appreciably like that of the Cordilleran geosyncline. The highly generalized geologic map of the Japanese Archipelago and the adjacent mainland (Fig. 7) is presented to show the sources of the sediments and the depths of the seas into which the sediments drain.

Figure 8 depicts the two great assemblages of sediments in the Cordilleran geosyncline as they were deposited during Silurian time. The Silurian was selected for illustration because it seemed to give a pattern that more closely simulates that of the Japanese region. In other periods, either volcanism was not prominent opposite both the Klamath and the Alexander troughs at the same time, or the seas did not spread over the eastern trough. Deep-water sediments, such as must be accumulating in the Sea of Japan and the

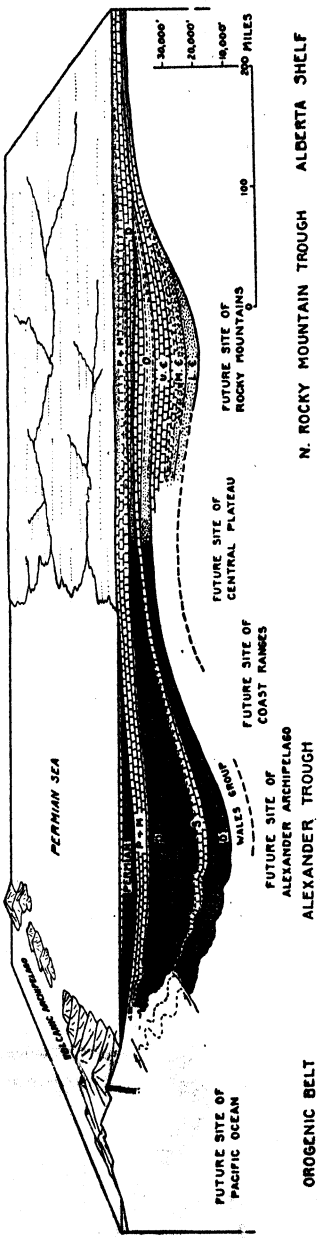


FIG. 9.—Cordilleran geosyncline in southeastern Alaska and British Columbia at the close of the Permian. The dark formations of the Alexander trough with white triangles are volcanics, and those with white triangles and round dots are graywacke. The other symbols are conventional.

Sea of Okhotsk, may be present in the Cordilleran geosyncline, but they have not been recognized.

A consideraton of the sedimentary conditions about volcanic archipelagos and the comparison of these with the mainland assemblage stimulate thoughts that ramify into many aspects of sedimentation but which are outside the objectives of this article. It must suffice to conclude that the distribution of sediments in the Cordilleran geosyncline indicates that a volcanic archipelago existed west of the western trough during the Paleozoic era.

ARCHIPELAGO AND OROGENIC BELT

The sedimentary evidence that establishes a great volcanic archipelago on the west side of the Cordilleran geosyncline at the same time indicates the existence there of an orogenic belt, because, from all we know, similar archipelagos in other parts of the world are the result of crustal movements in zones of active deformation.

Added to the sedimentary evidence of a Paleozoic orogenic belt is the dynamic evidence in the form of numerous unconformities between stratigraphic systems in the Pacific trough. Further dynamic evidence is the metamorphism of the trough's formations and, in addition to this, the convincing observation that the grains of the graywackes and the pebbles of the conglomerates were derived from already dynamically metamorphosed rocks of Paleozoic age. These, in short, are the lines of evidence that support the theory of a great orogenic belt to the west of the Cordilleran geosyncline during Paleozoic time. The theory is illustrated ideally in Figure 9.

CORRELATION OF THE EROSION SURFACES OF THE SOUTHERN APPALACHIANS

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ABSTRACT

Zonal profiles and field observations indicate that one peneplain, two erosion surfaces, and a minor surface or strath are present in the southern Appalachians, not considering present or recent valley floors. These surfaces are correlated with the well-recognized erosion surfaces of the northern Appalachians. The effect of structure upon the development of the erosion surfaces of the southern Appalachians is considered.

INTRODUCTION

The existence of erosion surfaces in the southern Appalachian region has been demonstrated by many workers.[1] Although these surfaces have been described in detail and named, little attention has been given to their correlation with the well-established surfaces of the northern Appalachian region. The purpose of this study is to establish a correlation between the erosion surfaces of the southern and northern Appalachian areas.

The index map (Fig. 1) shows the area of detailed study. The area chosen has the advantage of containing in the eastern part the folded structures of the Valley and Ridge and in the western part the broad Cumberland Plateau, where the structure is relatively simple and interpretation is easier than in the folded region. The nine quadrangles to the south are mapped in folios of the *Geologic Atlas of the United States*.

As early as 1890, Davis[2] suggested that the erosion surfaces of the northern Appalachians continue southwestward over the entire Appalachian region. In 1894, Hayes and Campbell[3] published an excellent and comprehensive paper on the geomorphic development of the southern Appalachians, in which they recognized two peneplains. Other students have already been mentioned, and the erosion surfaces recognized by these workers and the names applied to them are summarized in Table 1. The names are arranged in descending order from highest (oldest) to lowest (youngest) and are arranged to suggest correlations, where this is possible.

Recently, W. S. Cole[4] showed convincingly that in the northern Appa-

[1] W. M. Davis, "The Geologic Dates of Origin of Certain Topographic Forms on the Atlantic Slope of the United States," *Bull. Geol. Soc. Amer.*, Vol. II (1890), pp. 545–81; C. W. Hayes, "The Rome, Georgia-Alabama Folio," *Geol. Atlas of the United States* (1900), No. 78; C. W. Hayes and M. R. Campbell, "Geomorphology of the Southern Appalachians," *Nat. Geog. Mag.*, Vol. VI (1894), pp. 63–126; A. Keith, "Some Stages of Appalachian Erosion," *Bull. Geol. Soc. Amer.*, Vol. VII (1895), pp. 519–25; G. W. Stose and H. D. Miser, "Manganese Deposits of Western Virginia," *Va. Geol. Surv. Bull. 23* (1922), pp. 1–200; F. J. Wright, "The Physiography of the Upper James River Basin in Virginia," *Va. Geol. Surv. Bull. 11* (1925); "Erosional History of the Blue Ridge," *Jour. Sci. Lab. Denn. Univ.*, Vol. XXIII (1928), pp. 321–44; "The Older Appalachians of the South," *ibid*, Vol. XXVI (1931), pp. 143–250; "The Newer Appalachians of the South," *ibid.*; Vol. XXIX (1934), pp. 1–107; K. Ver Steeg, "Erosion Surfaces of the Appalachians," *Pan.-Amer. Geol.*, Vol. LVI (1931), pp. 267–84; "Correlation of Appalachian Peneplanes," *ibid.*, Vol. LXXIII (1940), pp. 203–10.

[2] Ftn. 1.

[3] Ftn. 1.

[4] "Nomenclature and Correlation of Appalachian Erosion Surfaces," *Jour. Geol.*, Vol. XLIX (1941) pp. 129–48.

lachians there is one peneplain, two erosion surfaces, and one strath stage. These surfaces are recognizable over the entire northern Appalachian district. Although local names are used for convenience in designating various surfaces, it was shown that satisfactory correlations can be made in the northern portion of the Appalachians.

Detailed study of the selected area in the southern Appalachians indicates that the same number of surfaces is present in both the northern and the southern Appalachians. Through the use of projected profiles, connecting the area of detailed study in the south with the area of demonstrated surfaces in the north, a correlation was established. This correlation is shown in Table 2.

METHODS OF STUDY

Two types of projected profiles were used in this study. For the district extending from the southern boundary of the Sewanee, Chattanooga, and Cleveland quadrangles to the middle of the Wartburg, Briceville, and Maynardville quadrangles, east-west lines were drawn across the maps at 1-inch intervals, and a zonal profile was made for each approximately 2-mile distance between the lines. The zonal profile was drawn on Bristol board, cut out, and mounted in a frame (Pl. I). For the Beattyville–Maynardville quadrangles a simple projected profile was made, the base line of which was along the western boundary of the quadrangles (Fig. 2).

The zonal-profile type has the advan-

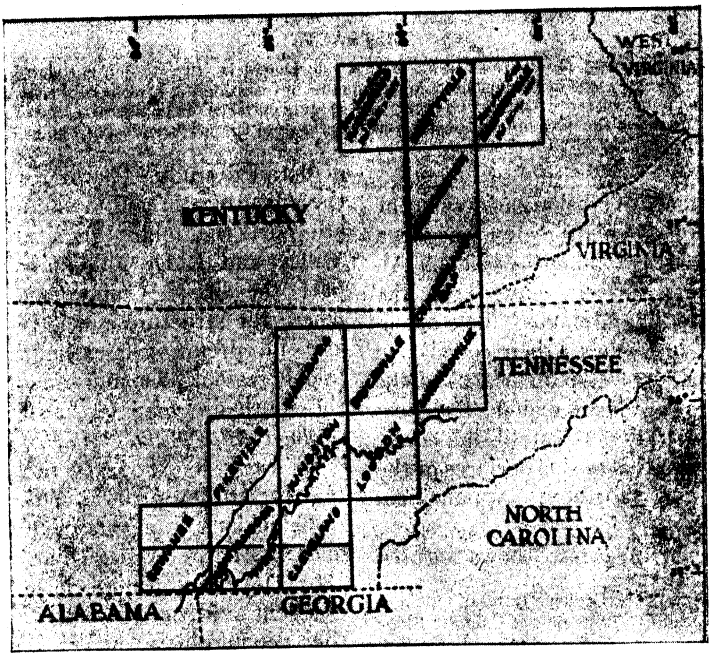
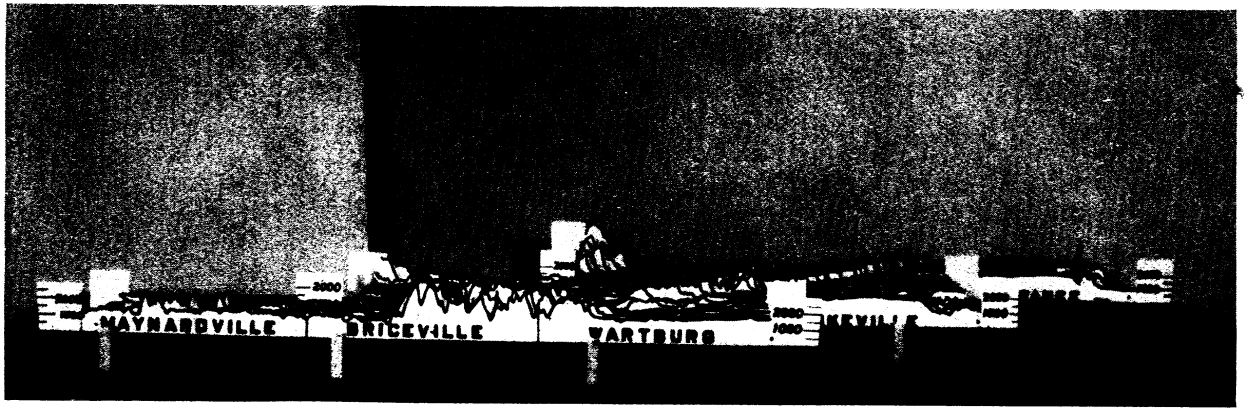
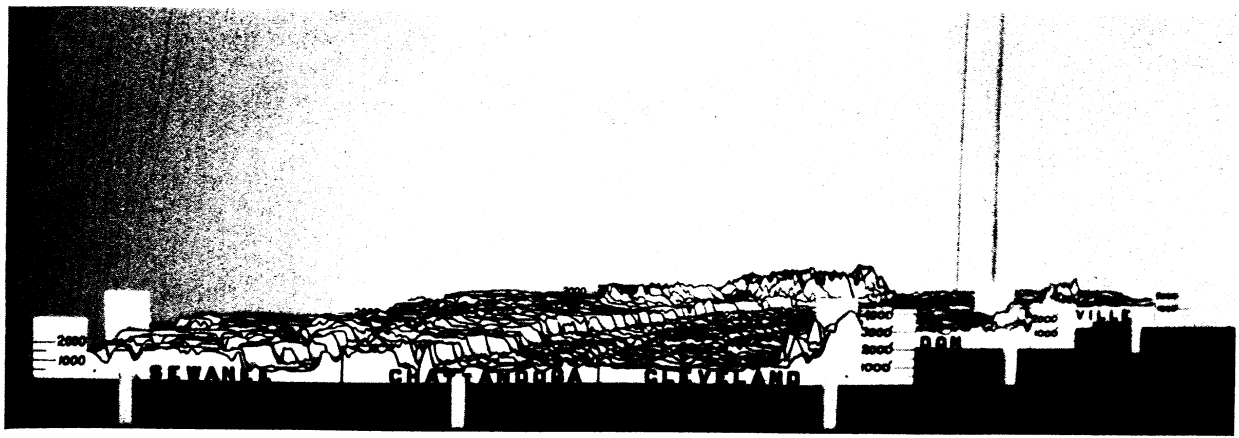


FIG. 1.—Index map, showing the area of detailed study in Tennessee and the quadrangles studied in Virginia and Kentucky. The heavy line on the west border of the Beattyville-Maynardville quadrangles was the base line for the projected profiles of this region. The east-west line through the center of the Sewanee-Cleveland quadrangles indicates the position of the structural section.

PLATE I



A, The model from north to south



B, The model from south to north

TABLE 1

Hayes and Campbell (1894)	Keith (1895)	Hayes (1900)	Stose and Miser (1922)	Wright (1928)	Ver Steeg (1931)	Wright (1931)	Wright (1934)	Ver Steeg (1940)	Shaffer (This Article)
Cretaceous peneplain	3,700′–3,000′	Cretaceous peneplain, 1,600′ in northern portion of valley region on Rome quadrangle	Summit peneplain, 4,000′–3,500′	Blue Ridge Summit Peneplain = Upland Peneplain	Schooley (Kitattiny)	Upland = Schooley	Schooley = Upland	Schooley	Upland
	2,600′–2,000′		Upland peneplain, 3,000′–2,700′						Cumberland = Allegheny
	1,800′–1,500′		Intermediate peneplain, 2,200′–2,300′						
Tertiary peneplain	1,100′–1,000′	Tertiary peneplain	Valley floor peneplains	Appalachian Valley Peneplain = Asheville = Piedmont	Harrisburg	Harrisburg (Valley)	Harrisburg (Valley)	Harrisburg = Lexington (Worthington)	Valley = Lexington (Worthington)
	Partial cycle represented by terraces and bottoms at 600′–700′	Coosa peneplain, 630′ in Coosa Valley	Terraces and benches along major transverse streams	Young to mature valleys of present cycle	Coosa = Somerville = possibly Lexington (Worthington)	? ? ? Coosa = Somerville	Rock terraces veneered with stream gravels and boulders	Parker strath	Coosa = Somerville = Parker strath

tage of giving a regional picture in three dimensions when all the individual zonal profiles are drawn on Bristol board, cut out, and mounted in proper sequence in a frame. These cutouts have another advantage. They may be removed from the mounting frame, compressed until they resemble a profile of the type used in the Beattyville–Maynardville section, and photographed and studied from both directions and from above (Pl. II).

TABLE 2

Northern Appalachians		Southern Appalachians
Western Portion	Eastern Portion	
Upland	Upland	Upland
Allegheny	Schooley	Cumberland
Lexington (Worthington)	Harrisburg	Valley
Parker strath	Somerville	Coosa strath

To demonstrate the influence of structure and stratigraphy on the development of the erosion surfaces, an east-west structural section was made, from the folios, through the middle of the Sewanee, Chattanooga, and Cleveland quadrangles (Fig. 3). The impressions gained by study of the profiles and structure sections were checked during three summers in the field.

THE MODEL

A regional picture in three-dimensional form is shown by the cutout profiles, mounted in a frame in proper sequence to form the model. The erosion surfaces and their relationship to the significant physical features of the southern Appalachians are best seen when the model is viewed from the south (Pl. I, *B*).

The highest surface, the Upland peneplain, preserved in the Crab Orchard Mountains, may be seen in the central background. After the Upland peneplain was formed near sea level, uplift occurred, and the Cumberland erosion surface developed.

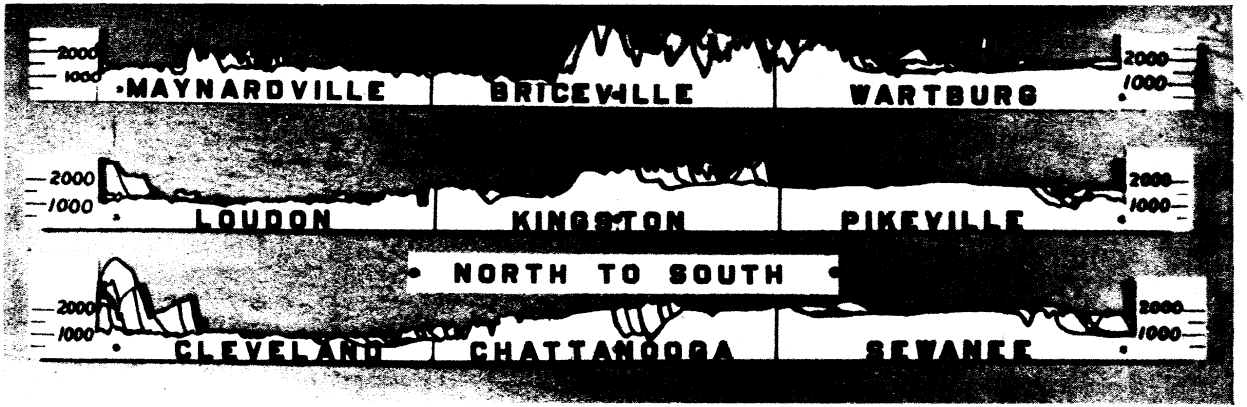
The Cumberland surface, the best-developed surface of the southern Appalachians, forms the top of the Cumberland Plateau and Walden Ridge Plateau and extends from the western border of the photograph eastward to the sharp topographic break approximately through the center of the photograph. This break, the Cumberland Escarpment, separates the Cumberland erosion surface above from the Valley erosion surface below.

Following uplift of the Cumberland surface, the Valley erosion surface, which increases in altitude northward, formed in the Valley and Ridge Province, shown in the eastern portion of the photograph. This surface has representatives in the hills in the Sequatchie Valley just east of the final *E* of "Sewanee" in the western foreground of the photograph.

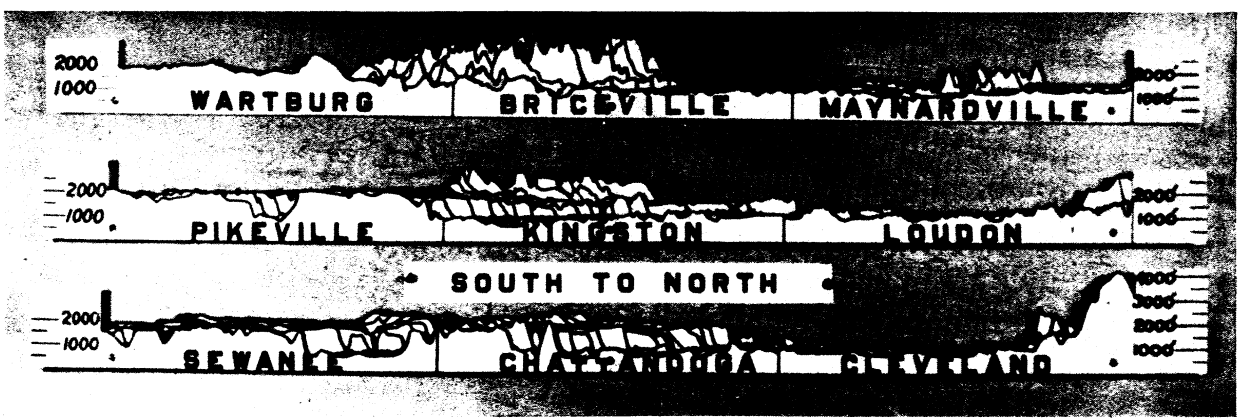
Again, after uplift of the southern Appalachians, the Coosa strath developed on weak rock along the major streams of the area. This surface is not well shown in the photograph because of mechanical difficulties inherent in projected profiles, but it may be seen in the field as gravel-covered stream terraces, about 650–700 feet above sea level. Slight uplift interrupted the extension of the Coosa surface, which had reached only the strath stage.

BEATTYVILLE–MAYNARDVILLE SECTION

The surface which extends completely across the Beattyville quadrangle at 1,300–1,400 feet and across the northern third of the Manchester quadrangle at 1,400–1,500 feet (Fig. 2) is the Lexington (Worthington) surface, which W. S.



A, Compressed profiles from north to south



B, Compressed profiles from south to north

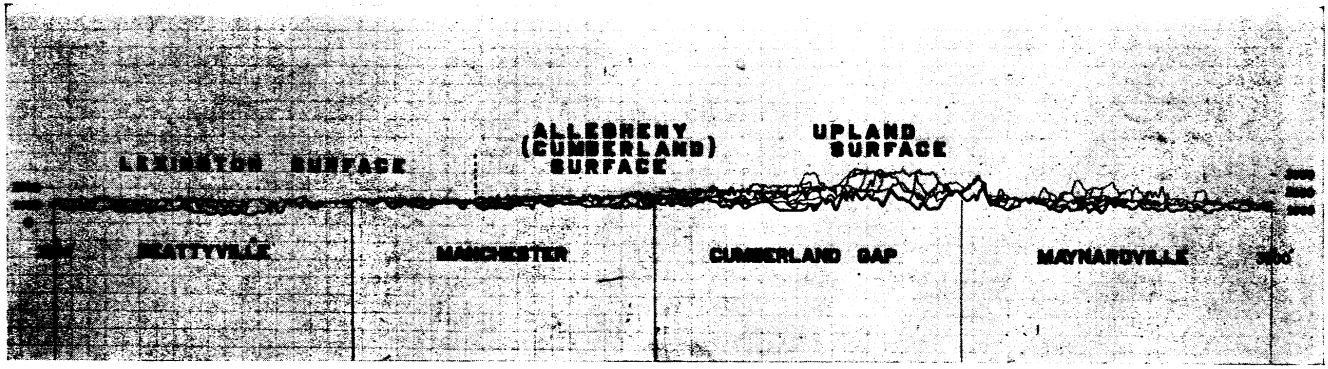


FIG. 2.—Projected profile of the Beattyville-Maynardville section. Base line along western boundary of the quadrangles. Vertical exaggeration approximately × 10.

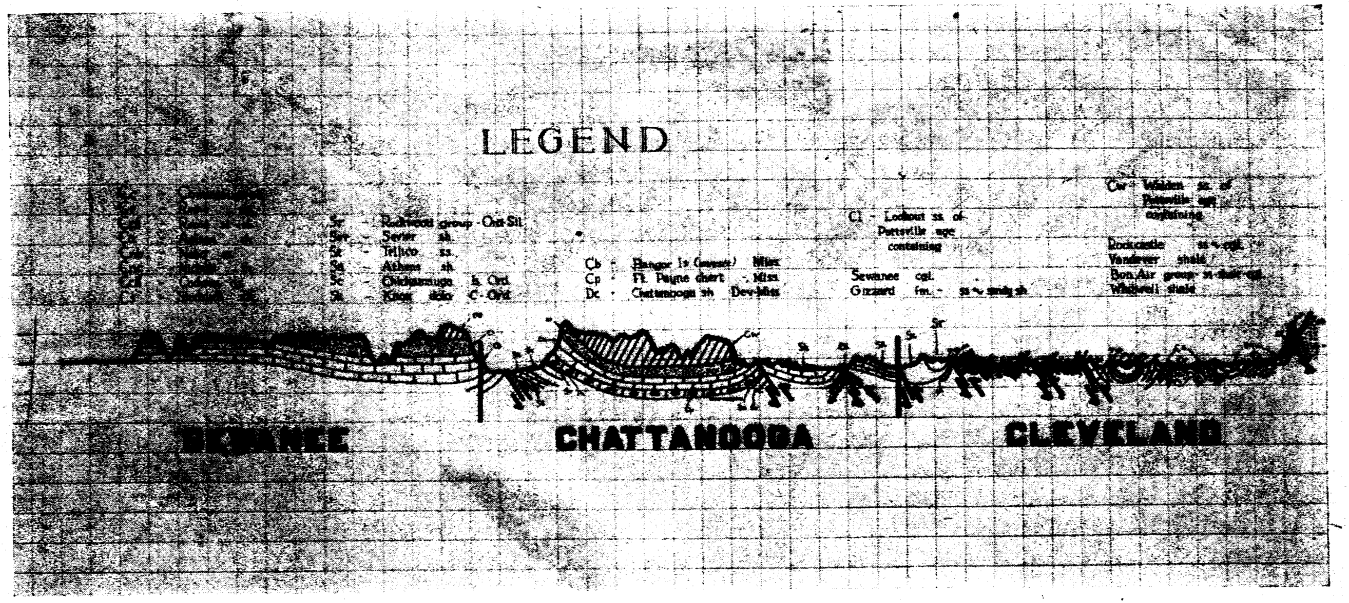


FIG. 3.—Structural section of the Sewanee-Cleveland area. Section drawn east-west through the center of the quadrangles. Vertical exaggeration approximately × 10.

Cole[5] described on the Richmond, Beattyville, and Salyersville quadrangles in Kentucky. There appears to be general agreement about this surface among workers in the plateau region of Kentucky and Ohio.

At approximately a third of the distance across the Manchester quadrangle a distinct break is apparent. This break is interpreted as the juncture of the Lexington (Worthington) and the Cumberland (Allegheny) surfaces. The latter is the surface which extends across the Cumberland Plateau farther south and was called "Cumberland" by Hayes.[6] On the quadrangles to the south the parenthetical "Allegheny" will be omitted. The break just noted has been identified throughout the tract from central New York southward to the area of this study. It has been described by W. S. Cole[7] on the projected profiles of the New Martinsville quadrangle (1934), the Salyersville, Wayne, and Arnoldsburg quadrangles (1935), the Centerpoint, Freeport, and Shenango quadrangles (1937), and the Cambridge Springs and Honeoye quadrangles (1938). The geologic map of Kentucky shows no stratigraphic reason for the break, and it is therefore interpreted as an erosional feature.

At its juncture with the Lexington (Worthington) surface the Cumberland surface is at an elevation of 1,600 feet and rises southward on the Manchester quadrangle until it reaches 1,900 feet at the southern border. A long strath of the Lexington (Worthington) surface extends into the Cumberland surface in the center of the Manchester quadrangle at an elevation of 1,500 feet. The Cumberland surface continues across the northern third of the Cumberland Gap quadrangle at 1,900–2,000 feet. Southward some reduced portions of the Upland peneplain are present at elevations of 2,300–2,400 feet. These are followed southward by summit areas of the Log Mountains and Cumberland Mountain at 3,000–3,300 feet above sea level. At the juncture of the Cumberland surface and reduced remnants of the Upland peneplain, a long strath of the Cumberland surface, at an elevation of about 2,000 feet, extends into the reduced Upland remnants. The Upland peneplain, or reduced portions of it which are higher than the Cumberland surface, extends across the southern two-thirds of the Cumberland Gap quadrangle.

The projected profile of the Maynardville quadrangle overlaps with the cutouts made of the southern half of the same quadrangle. The interpretation is made with the regional picture in mind, especially the continuation of the surfaces to the south in the nine-quadrangle area of detailed study. The peneplain and two erosion surfaces are present; the minor strath stage is not shown because of mechanical difficulties present in projected profiles, but its existence is demonstrated by field studies. There are remnants of the Upland peneplain, represented by portions of Cumberland Mountain at altitudes of over 2,900 feet, a few peaks of Clinch Mountain at 2,700 feet, and some parts of Lone Mountain at 2,200–2,300 feet above sea level. There are a very few remnants of the Cumberland surface at altitudes of from 1,900 to 2,100 feet. This surface is poorly pre-

[5] "Rock Resistance and Peneplain Expression," *Jour. Geol.*, Vol. XLIII, Part II, No. 8 (1935), p. 1056.

[6] P. 2 of ftn. 1.

[7] "Erosion Surfaces in Eastern and Southern Ohio," *Jour. Geol.*, Vol. XLII (1934), p. 291; p. 1050 of ftn. 5; "Development and Structural Control of Erosion Surfaces," *ibid.*, Vol. XLV (1937), p. 145; "Erosion Surfaces of Western and Central New York," *ibid.*, Vol. XLVI (1938), p. 193.

served on the Maynardville quadrangle and in the Valley and Ridge generally. The Valley surface, named by Wright,[8] is present at an elevation of 1,500–1,600 feet. It is well shown on the southern fourth of the quadrangle and extends northward as a strath below the remnants representing the two higher surfaces.

WARTBURG–MAYNARDVILLE SECTION

In the Wartburg–Maynardville section, as seen from the south (Pl. II, *B*), the Upland peneplain and two erosion surfaces are shown. In the western half of the Wartburg quadrangle the Cumberland surface, forming the top of Cumberland Plateau, is well shown at 1,800–2,000 feet above sea level. Just east of the central part of the Wartburg quadrangle a peak of the Crab Orchard Mountains, representing a reduced portion of the Upland peneplain, rises to an elevation of 2,500 feet. The Upland is present in the eastern part of the Wartburg quadrangle in the Crab Orchard Mountains at elevations of 2,500 to approximately 3,000 feet. When viewed from the north (Pl. II, *A*), the Cumberland surface may be seen continuing below the higher elevations of the reduced Upland peneplain.

On the Briceville quadrangle the Upland peneplain and two erosion surfaces are present. The surfaces are best shown when the profile is viewed from the south (Pl. II, *B*). The crests of the Crab Orchard Mountains at 3,200–3,500 feet above sea level in the western and central parts of the quadrangle represent the Upland peneplain. In the central portion of the quadrangle below the crests of the Crab Orchard Mountains but above the Valley surface there is a strath of the Cumberland surface at 1,700–1,800 feet. Below the Cumberland surface the Valley surface is well shown in the eastern portion of the quadrangle at elevations of 1,200–1,300 feet. A few prominences rise above the Valley surface to an elevation of 1,500 feet. In the central portion of the Briceville quadrangle the Valley surface extends westward as a strath below the level of the Cumberland surface.

The Valley surface, best seen from the south (Pl. II, *B*), extends across the Maynardville quadrangle at 1,600–1,700 feet in altitude. It is present as a strath below the few peaks of Clinch Mountain, with elevations of 2,200–2,500 feet. The higher portions of these peaks represent greatly reduced remnants of the Upland peneplain, and a few peaks at elevations of 1,800–2,000 feet represent the very poorly preserved Cumberland surface.

PIKEVILLE–LOUDON SECTION

On the profiles of the Pikeville quadrangle, best shown when viewed from the south (Pl. II, *B*), the Cumberland surface forms the top of Cumberland Plateau at elevations of 2,000–2,100 feet west of the Sequatchie Valley. The same surface continues across the top of Walden Ridge to the east of Sequatchie Valley, at elevations of 2,000–2,100 feet. Within the Sequatchie Valley, in the central portion of the quadrangle, is a group of hilltops with elevations of about 1,200 feet, which are interpreted as representing the Valley erosion surface.

The profile of the Kingston quadrangle shows the peneplain and the two erosion surfaces. The Cumberland surface extends completely across the quadrangle at elevations of from 1,800 to 2,000 feet. To the west it is present as a strath below the Upland peneplain, and in the east it stands out above the Valley surface. The Valley surface extends completely across

[8] P. 335 of ftn. 1 (1928).

the quadrangle at elevations of from 1,100 to 1,300 feet. Although the Coosa strath, named by Hayes,[9] is difficult to identify on profiles of this type, the flat area between 600 and 700 feet above sea level in the western portion of the Kingston quadrangle may represent this strath.

With the exception of a few ridge-top elevations in the western portion of the quadrangle, which represent monadnocks on the Valley surface, and a few peaks in the extreme eastern portion of the quadrangle, composed of resistant material, the Valley surface extends across the Loudon quadrangle at elevations of 1,100–1,300 feet. This surface is seen to best advantage when the profile is viewed from the north (Pl. II, *A*).

SEWANEE–CLEVELAND SECTION

The Cumberland surface extends across the western three-fourths of the Sewanee quadrangle at elevations of 1,900–2,100 feet. In the eastern quarter of the quadrangle there are some reduced portions of the Upland peneplain at elevations of 2,300–2,400 feet. The hilltops in the Sequatchie Valley at elevations of 1,000–1,100 feet are interpreted as representatives of the Valley surface.

In the western portion of the Chattanooga quadrangle some summit areas rise over 2,400 feet above sea level, especially on the eastern rim of Walden Ridge, overlooking the Sequatchie Valley. These higher areas represent reduced portions of the Upland peneplain, which is better developed farther north. The Cumberland surface extends almost completely across the quadrangle at elevations of 1,900–2,100 feet. In the western part of the Chattanooga quadrangle it may be seen as a strath beneath the higher peneplain. The Valley surface stands out clearly at 1,000–1,200 feet in the central part of the quadrangle as a surface below the Cumberland when seen from the south (Pl. II, *B*).

Below a few scattered peaks of White Oak Mountain in the western part of the Cleveland quadrangle the Valley surface is developed at elevations of 1,000–1,100 feet. This is best shown when the profile is viewed from the north (Pl. II, *A*). To the east the Valley surface continues completely across the quadrangle below the peaks of Beans Mountain and Big Frog Mountain. The latter has peaks rising over 4,300 feet.

The Coosa surface, which is a minor surface or strath, has been mentioned by practically every worker in the southern Appalachians. It can be seen in the field in the form of terraces covered with stream gravels along the major streams and on some of their tributaries between 650 and 700 feet above sea level.

STRATIGRAPHY AND THE EROSION SURFACES

The Upland peneplain is best developed in the highest portions of the Crab Orchard Mountains, which are capped by resistant Pottsville rocks. The Crab Orchard anticline, as pointed out by W. A. Nelson,[10] is the partially eroded extension of the Sequatchie Valley anticline. Nelson states: "The higher mountains occur on the western and the southern ends. On the top of these ridges and in the old wind gaps the strata have dips of 25 to 34 degrees, but on the mountains that lie on the axis of the anticline the strata lie practically flat or have dips of 5 to 10 degrees to the northwest or southeast."

The Cumberland erosion surface is splendidly developed on top of Cumber-

[9] P. 2 of ftn. 1.

[10] "The Southern Tennessee Coal Field," *Tenn. Div. of Geol., Bull. 33-A* (1925), p. 57.

land Plateau and on Walden Ridge. The cap rock consists of sandstones and conglomerates of Pottsville age, which are strongly resistant to weathering and erosion. The Cumberland surface, however, truncates the structures and the rock units, regardless of their resistance (Fig. 3). Nelson[11] points out that, although the structure underlying the Cumberland Plateau is, in general, horizontal, the area is covered by low anticlines, synclines, and domes, which follow one another in confusing succession. The most perfect erosion surface in the area, the Cumberland surface, truncates these structures and the rock units.

The Valley erosion surface is developed on a very complex set of closely folded and faulted anticlines and synclines in the Valley and Ridge Province. The surface is developed on weak rock units, both limestones and shales, and extends up some of the major valleys in the area, especially into the Sequatchie Valley.

The Coosa surface is a strath in the form of gravel-veneered terraces along the major streams of the area. It is developed on weak rock, principally limestone.

CONCLUSIONS

One peneplain, two erosion surfaces, and a minor surface or strath are present in the southern Appalachians, if the present or recent valley floors are excluded. These surfaces from the oldest (highest) to the youngest (lowest) are: (1) Upland peneplain, (2) Cumberland erosion surface, (3) Valley erosion surface, and (4) Coosa strath.

The term "Upland" is used, as it was employed by Cole[12] in the northern Appalachians, for a peneplain which is higher than the Allegheny erosion surface of the northern Appalachians and above the Cumberland erosion surface of the southern Appalachians.

The first erosion surface below the Upland peneplain is called the "Cumberland erosion surface," and the term is used as a correlative of the Allegheny surface of the northern Appalachians, as used by N. M. Fridley and J. P. Nölting, Jr.[13] It is well developed at an altitude of 1,800–2,100 feet on top of the Cumberland Plateau, on resistant strata of Pottsville age.

Below the Cumberland occurs the Valley erosion surface, which occupies the same relative position as the Lexington (Worthington) surface in the northern Appalachians and is regarded as a correlative of it. It is best developed at elevations of from 1,000 to 1,300 feet in the Valley and Ridge Province on a very complex system of closely folded and faulted anticlines and synclines. It is preserved on weak rock, limestone, and shale. The Valley surface rises gradually northward toward the divide area in the New River region.

The Coosa strath, an incomplete surface, is the lowest of the series and is preserved on weak rock along the major streams. The Coosa is regarded as a correlative of the Somerville of New Jersey and the Parker strath of the northern Appalachians.

ACKNOWLEDGMENTS.—The writer gratefully acknowledges his indebtedness to Dr. W. Storrs Cole, of Cornell University, and Dr. P. B. Stockdale, of the University of Tennessee, for their aid during the field work and the preparation of this paper. The Graduate School of Ohio State University granted financial assistance for the field studies and the photography.

[11] *Ibid.*, p. 54.

[12] P. 135 of ftn. 4.

[13] "Peneplains of the Appalachian Plateau," *Jour. Geol.*, Vol. XXXIX (1931), p. 753.

THE INFLUENCE OF GEOLOGIC STRUCTURE ON THE DRAINAGE PATTERN IN NORTHEASTERN MINNESOTA

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ABSTRACT

The region is especially interesting from the standpoint of geomorphology. Erosion of monoclinal structure of the Animikie and Keweenawan formations has developed ridges with an asymmetrical profile, known as "sawtooth mountains." The ridge and valley topography in the area underlain by the Rove slate and associated intrusives (dikes and sills) resembles that of the folded Appalachians. The preglacial drainage pattern was trellised.

The Duluth gabbro is banded; those zones composed of minerals less resistant to weathering and erosive agencies underlie the valleys, and the more resistant rocks, such as the "red rock," make up the ridges.

The lakes located on the Saganaga granite are irregular in outline and are not oriented in any particular direction.

The Pleistocene ice sheets produced noteworthy changes in the drainage lines. Numerous glacial lakes now occupy depressions which show striking east-west alignment. Many of the streams are new or have been forced to find new channels. Waterfalls and rapids and postglacial gorges are common.

INTRODUCTION

The features described in this paper lie for the most part within the Gunflint district of the Superior National Forest in northeastern Minnesota, extending from Lake Superior on the south to the Canadian border on the north and east and to the Cook County line on the west.

A study of the map of the area reveals, almost at first glance, a pronounced orientation of the lakes in an east-west direction. This is especially true of the lakes located east of a line running from the western end of Gunflint Lake southward to Brule Lake and Lake Superior. Owing to the banded character of the Duluth gabbro, there is some degree of alignment west of that line. In the region underlain by homogeneous granite and granite gneiss, north of Gunflint Lake, North Lake, and Arrow Lake, there is no pronounced east-west alignment.

It is the purpose of the writer to show the relation of structure and rock composition to the surface features and to determine the late Tertiary drainage for part of the area.

GENERAL SURFACE FEATURES

The area lies within the physiographic province known as the "Superior upland." It is underlain by coarse-grained, homogeneous igneous rocks of Archean and Algonkian age and by banded, alternating weak and resistant, tilted rocks, mainly sediments and lavas of Proterozoic age. The area is an upland cut by broad, open valleys and diversified by monadnocks, which consist mainly of elongate ridges and rounded higher masses. One who has a preconceived notion that the Laurentian upland, of which the region is a part, is a gently rolling surface of little relief, is due for a rude awakening when he views this area with its rough topography. A drive over the Gunflint Trail from Grand Marais northward to Gunflint Lake is over a seemingly endless series of hogbacks or ridges, giving a gigantic washboard effect. These ridges end where the homogeneous Saganaga granite makes its appearance.

In an area of complex structure and diversified rock composition, an erosion

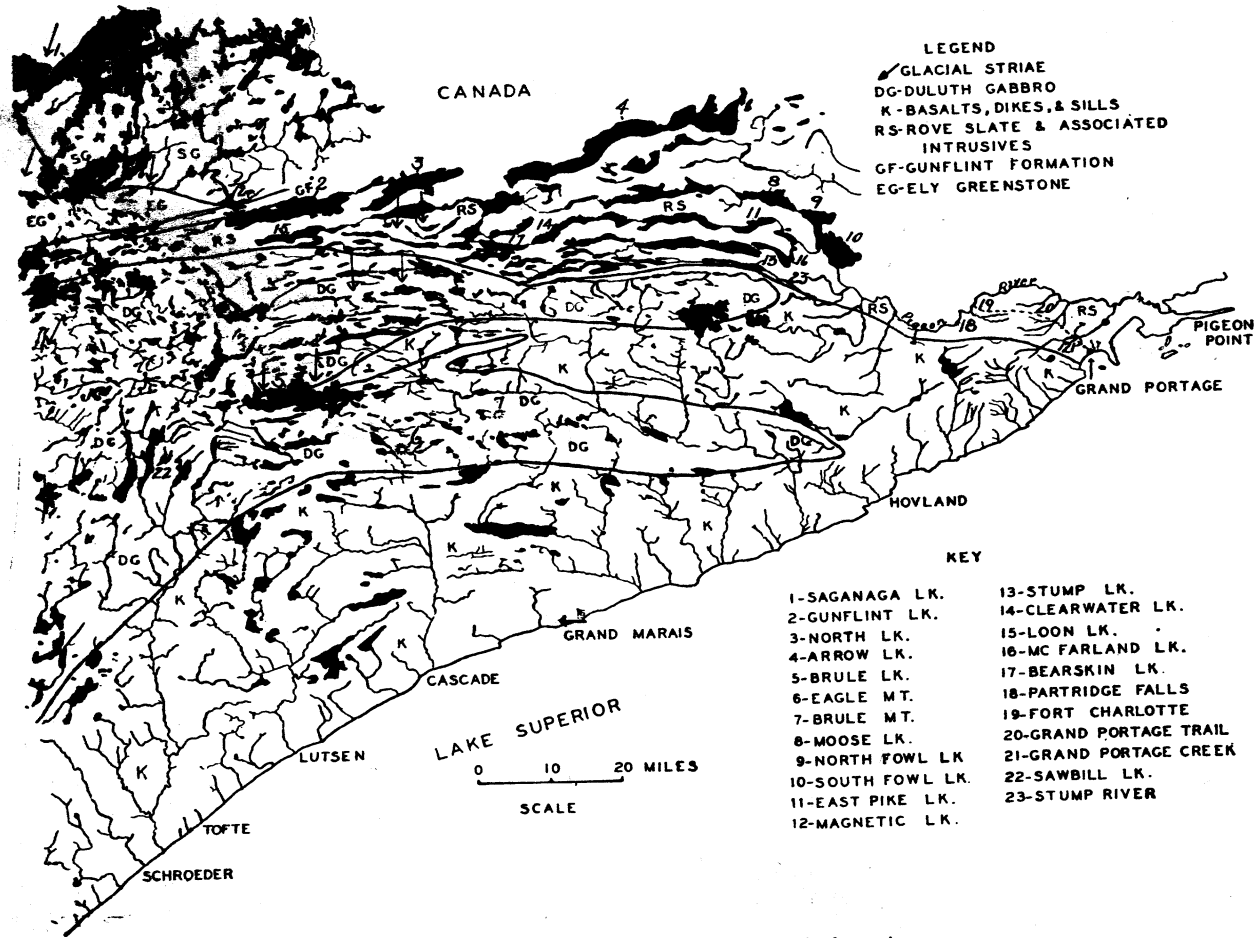


FIG. 1.—Map of area, showing the distribution of the formations

surface may exhibit considerable variation in elevation, being lower on the weaker rocks and higher on the more resistant ones. There is always danger of misinterpreting one erosion surface for many in such a region.

Where the homogeneous, coarse-grained granites, greenstones, schists, and gneisses of Archean age are present, the rolling peneplain shows up to good advantage. In the area of Proterozoic rocks, the Animikian series is composed of the resistant Gunflint iron-bearing formation and the weaker Rove slate. The Keweenawan series includes a great thickness of basalt lava flows, interstratified with sedimentary beds and dipping gently toward Lake Superior to the south.

The Logan diabase sills and dikes intrude the Keweenawan flows and older formations. They dip to the south and form ridges where they intruded into weaker rocks, such as the Rove slate. In the western and central portion of the area, the Duluth gabbro lopolith, which was intruded at the base of the Keweenawan flows, forms a broad upland. The banded composition of the gabbro results in a series of minor ridges standing above the peneplain, not so conspicuous as those in the Rove slate region but of considerable relief. They commonly trend in a slightly northeast-southwest or east-west direction. Rounded knobs of more resistant rock, such as the "red rock" in the area of the Duluth gabbro, are perhaps the most conspicuous of the monadnock features. None is of great extent or of great height, and none would be noticeable were it not for the gently rolling character of the surrounding region, although, in general, they stand higher than the numerous ridges in the area. Misquah Hill (2,230 feet), in Cook County, is the highest point in Minnesota. Among other conspicuous monadnocks are Eagle Mountain and Brule Mountain.

The skyline appears to be horizontal, most elevations forming a common level at about 1,600–1,700 feet. Many points, 1,700–1,800 feet and higher, stand above the general level, and a few monadnocks reach 2,100–2,300 feet. The long, gentle slopes of the broad, open valleys between the ridges lie between 1,300 and 1,600 feet, the average for the areas on the weaker rocks being about 1,300–1,400 feet. The maximum relief of the peneplain is perhaps 400–500 feet, and the average about 100–200 feet.

The Laurentian peneplain was probably developed in the long period of crustal stability in Mesozoic time. If this theory is correct, the peneplain near the summit of the Laurentian upland is of the same age as the Schooley (Kittatinny) erosion surface, widely developed throughout the whole of the Appalachian and New England provinces. Considerable areas within the Laurentian upland may be parts of the resurrected pre-Cambrian erosion surface, which has been little modified since it was uncovered by erosion. After the development of the summit peneplain and long before the glacial epoch, the region was uplifted. The rejuvenated streams dissected the uplands, and at least one lower erosion surface and possibly more were developed. The upland was deeply dissected, broad valleys were developed, and a complex network of drainage lines was established before the advent of the ice sheets.

Outstanding among the physiographic features of the area are the asymmetrical ridges with the long dip-slope to the south and the steeper northward-facing scarps. A view from the highway, east of Grand Marais and at many other points,

shows a skyline with a sawtooth profile. This is especially true of the hills in the vicinity of Lutsen and Tofte. When viewed at an angle, a single ridge broken by gaps may give the false impression of a series of asymmetrical ridges. A number of ridges in succession give the same effect when viewed from a distance. Hence these hills are referred to as "sawtooth mountains."

STRUCTURAL FEATURES

The general structural trend of the rocks in the region, north of Lake Superior, is about N. 70° E., the alignment of the ridges corresponding with the major structures. The structural alignment in the region of the Rove slate is that of a great curve, with the ridges bending slightly to the southeast, in the vicinity of Moose Lake, North Fowl Lake, South Fowl Lake, Pike Lakes, and McFarland Lake. From there eastward, the ridges bend to the northwest in the area north and northwest of Grand Portage.

The most important structural characteristic of the area is the general dip of the formations to the south toward Lake Superior. Where the dipping beds are of unequal resistance to erosion, they produce hogbacks or cuestas, depending on the degree of inclination. On the rocks of homogeneous character, such as granites, there is no alignment of topographic features or drainage.

ELY GREENSTONE (KEEWATIN)

The Ely greenstone underlies a small section of the area and outcrops a few miles to the west of the narrows between Gunflint and Magnetic lakes. It is largely a chlorite schist and shows structures characteristic of lava flows. The intrusion of the Saganaga granite has altered it to hornblende schist. In general, the Ely greenstone is resistant and forms broad, low, rounded knobs and ridges. Farther west, in the Ely district, it forms prominent hills and ridges trending east-west.

SAGANAGA GRANITE (LAURENTIAN)

The Saganaga granite is a typical batholithic mass and is exposed in an area 15 by 25 miles along the boundary of Minnesota and Ontario. It has been deeply eroded, and its eastern exposures are believed to have been many miles below the surface before erosion began.[1]

The granite is intrusive into the Ely greenstone, which it has altered near the contact. It is composed chiefly of orthoclase, plagioclase, quartz, biotite, and hornblende. The Saganaga Lake granite weathers readily, the abundant feldspar changing chemically to kaolin.

The lakes are numerous and very irregular in outline in the homogeneous granite and without definite alignment. The granite area stands somewhat lower than the Ely greenstone and Duluth gabbro to the south and has a gently rolling peneplaned surface of low relief.

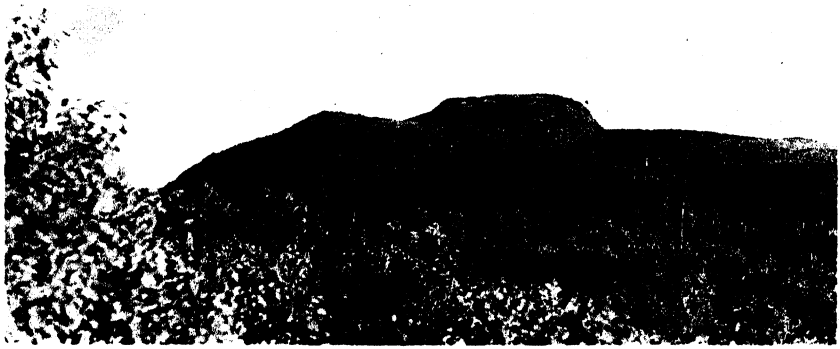
GUNFLINT IRON FORMATION (ANIMIKIE)

Monoclinal ridges are not restricted to the Keweenawan series. In the Huronian, diabases and quartzites have developed monoclinal ridges with an asymmetrical profile of gentle dip-slopes to the south and steeper faces to the north.

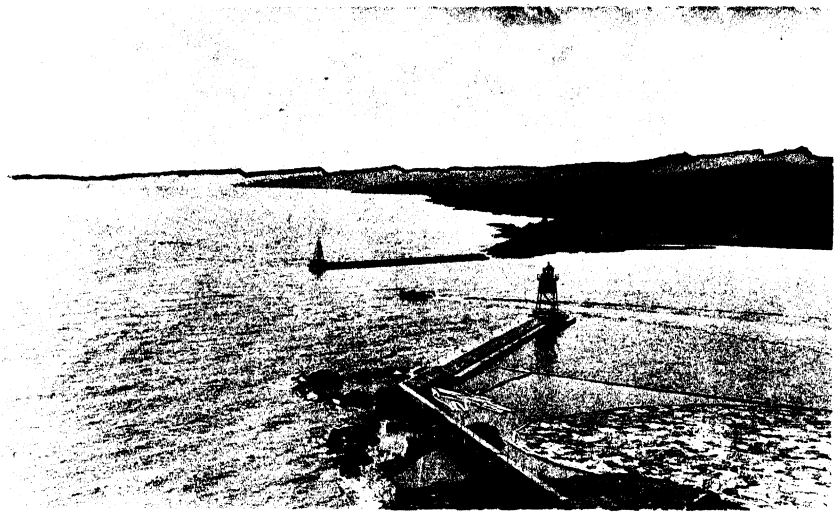
The Gunflint iron-bearing formation of Middle Huronian age is believed to be the equivalent of the Biwabik iron formation of the Mesabi Range. It is from 300 to 700 feet thick in the Gunflint area and may be observed on the north shore and

[1] F. F. Grout, "Structural Features of the Saganaga Granite of Minnesota-Ontario," *Rept. XVI Internat. Geol. Cong.* (Washington, 1933), pp. 255–70; "The Saganaga Granite of Minnesota-Ontario," *Jour. Geol.*, Vol. XXXVII (1929), pp. 562–91.

PLATE I



A



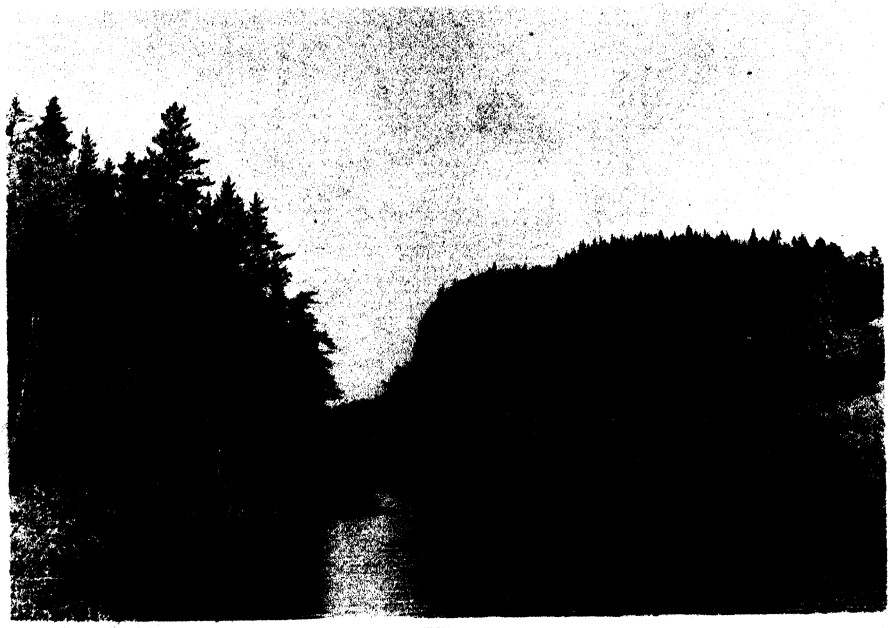
B

A, Carleton Peak, a glaciated monadnock of the Superior upland

B, a view west of the highway near Grand Marais. Note the sawtooth skyline of the Keweenawan basalt flows.

PLATE II

A

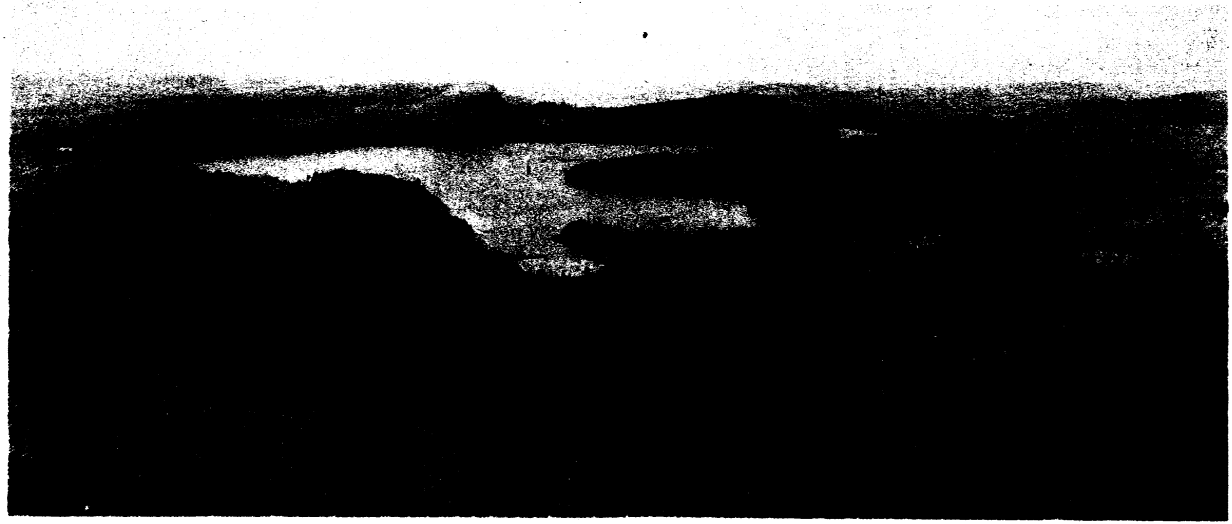


B

A, Sea Gull Lake, a typical lake in the Saganaga granite area

B, a typical ridge in the Rove slate area, showing the north-facing escarpment and talus and the dip-slope of a diabase sill. Mountain Lake in the foreground.

PLATE III



Steep north-facing cliffs and gentle southward dip-slopes of the diabase sills overlooking the Rove slate. Note the sawtooth topography of the diabase sills. McFarland Lake, looking west.

to the west of Gunflint Lake. Its outcrop belt trends in an almost east-west direction, parallel with other formations in the Animikie series, which widens in the form of a V to the eastward. The iron formation is exposed only in the vicinity of Gunflint Lake. Increasing amounts of the Rove slate are exposed to the eastward.

ROVE SLATE (UPPER HURONIAN) AND ASSOCIATED INTRUSIVES

The Rove slate received its name from Rove Lake, where it is well developed. The westernmost outcrop of the Rove slate is at the western end of Gunflint Lake; it widens rapidly to the east. The northern boundary of the slate extends northeastward and is limited by the Gunflint iron formation. The Duluth gabbro, which marks its southern boundary, trends east-southeast. The greatest width of the Rove slate in the United States is about 2 miles, and much of this is taken up by sills and dikes, which are intruded into it.

The thickness of the Rove varies, being thinner to the west and thicker to the east on the Canadian side of the boundary, where the larger part of the formation is exposed. A section from Moose Lake to Stump Lake is 3,000 feet thick, composed of about equal amounts of slate and diabase sills. In general, the beds strike about N. 70° E. and commonly have gentle southeast dips, varying from 4° to 15°. Steeper dips and minor folds are present near the Duluth gabbro and other intrusives. Joints are conspicuous and very regular in places. Faults of small throw are present but difficult to trace. They are of little importance in the formation of the conspicuous north-facing slopes.

The structure of the Rove slate is brought into relief by the intrusive dikes and sills, which Grout estimates at about half the volume of the formation.[2] The southward-dipping sills from Gunflint Lake to Stump Lake produce sawtooth mountains with summits regularly spaced. Other sills are present in the area east of Stump River to Pigeon River, but this topography is made more complex by many dikes. The Logan sills stand out as the most prominent topographic features of the region. The high ridges and especially the northward-facing bluffs are, for the most part, composed of intrusive diabase. Exposures are easily reached on the south side of Gunflint, Clearwater, Loon, West Bearskin, and McFarland lakes. Nearly all these lakes and others within the Rove slate area have a scarp or bluff of diabase on the south side. South of Clearwater Lake, for a distance of 3 miles, cliffs of diabase rise to an elevation of 2,000 feet above sea-level and a height of 280–320 feet above lake-level. Many of the northward-facing bluffs contain several sills, which alternate with slate and commonly have a thick sill at the crest.

In the vicinity of Grand Portage and from that locality north and east to Pigeon Point, dikes seem to be more common than sills, but both dikes and sills penetrate the Rove formation throughout the area in Minnesota. The dikes form ridges that are just as conspicuous as the sawtooth mountains formed by the sills, but the dike ridges are different, in that they have a bluff on both sides. The sills are more abundant in the slates, while the dikes are more common in the graywackes and the massive, jointed quartzites. Similar dikes and sills are present in the underlying Gunflint iron formation and in the overlying Keweenawan flows.

[2] F. F. Grout and G. M. Schwartz, "Rove Formation and Associated Intrusives in Northeastern Minnesota," *Bull. Minn. Geol. Surv.* (1933), p. 25.

The diabasic sills of the Rove formation vary from a few feet to more than 500 feet in thickness. A single series of outcrops, indicating a continuous sill, may show a greater thickness of diabase at some points than at others. Many small sills are concealed because the larger ones form prominent ridges, whose talus slopes cover not only the underlying slate but small sills intruded into the slate. Talus accumulations of angular blocks of diabase at the base of the bluffs in many places cover the contact between the diabase and the slate.

TOPOGRAPHY OF THE AREA UNDERLAIN BY THE ROVE SLATE AND ASSOCIATED INTRUSIVES

The topography resulting from the erosion of the weak Rove slate, with its dense diabase sills and dikes, is that of subparallel ridges and broad, open valleys, similar to that of the folded Appalachians, where alternating formations of weak and resistant rocks form parallel ridges and valleys. In the Rove slate area the ridges are not so high as in the Appalachian region; the dips of the beds are lower (5°–10°), and the structure is predominantly monoclinal. The Rove slate topography has developed on a huge monocline, which dips south toward the Lake Superior Basin and once formed the southern flank of a mountain range.

The valleys located on the Rove slate are now occupied by an east-west series of beautiful glacial lakes. The preglacial streams have been greatly modified by glaciation. Transverse valleys are not numerous, but some have important effects on the drainage. The ridges rise to as much as 2,100 feet above sea-level north of Clearwater Lake. Others rise to 2,000 feet, and many stand between 1,800 and 1,880 feet. Where they are highest, the ridges average about 1,850 feet. Farther east, nearer Grand Portage, the ridges have lower elevations, ranging from 1,400 to 1,720 feet. Where the Duluth gabbro upland region bounds the Rove area on the south, its elevations correspond to the highest ridges in the latter.

The lowlands on the Rove slate are at elevations varying from 895 to 1,850 feet, as indicated by the lake levels; but, as some of the lakes are deep, the valleys in which they lie must have been cut still lower. It is reported that many of the east-west–trending lakes are deepest near the south side, a condition to be expected with the existing monoclinal structure. The average elevation of the valley bottoms is not far from 1,650 feet; the average relief is about 200 feet. In general, the elevation of ridges and lowlands decreases as one approaches Grand Portage.

Figure 2 shows the drainage as it probably was in late Tertiary time. The pattern resembles that of the Appalachian Ridge and Valley Province. Windgaps and watergaps were present, although not so large or numerous as in the Appalachians. Doubtless, the drainage was thoroughly adjusted to the structure in the late stage of the cycle of erosion.

It is presumed that the preglacial streams in the parallel valleys on the weak slate drained eastward through the broad, open valley from South Fowl Lake and also along the wide lowland now occupied by Stump and Pigeon rivers, to a point west of Partridge Falls and thence continued in a southeasterly direction. Another possible route for the preglacial waters would be from the ancient valley above Partridge Falls, following the valley eastward from the site of Fort Charlotte along the route now followed by the Grand Portage trail (Fig. 3). The broad open gaps in the dia-

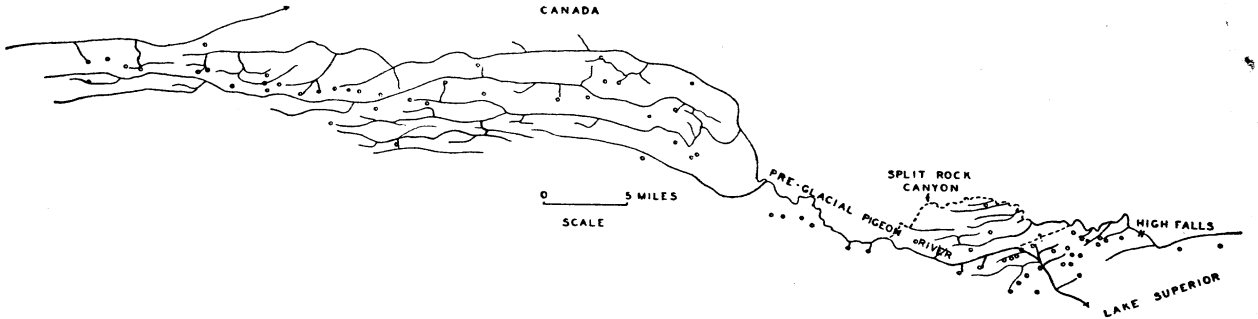


FIG. 2.—Diagram showing preglacial drainage pattern in the Rove slate area. Dashed line represents the portion of the Pigeon River which shifted as a result of glaciation. The circles indicate the location of gaps.

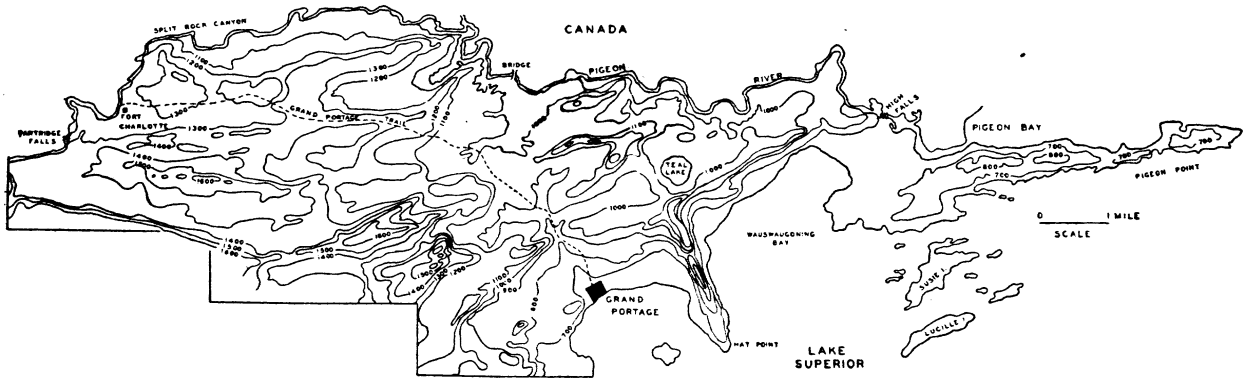


FIG. 3.—Map showing the topography in the vicinity of the Grand Portage

base ridges back of Grand Portage are now occupied by a small stream—Grand Portage Creek—and were doubtless cut in preglacial time by a river considerably larger than the creek now passing through them. Whether the preglacial stream was one of small drainage area or carried the waters from an extensive land surface to the west is impossible to determine. The gaps in the ridges northwest of Grand Portage are ½ mile wide, and it is probable that a stream of considerable size cut them.

The youthful character and circuitous route of the Pigeon River, across numerous resistant diabase ridges instead of by a more direct course to Lake Superior, can be explained as a result of glaciation. It would seem that a more direct route, parallel to the diabase ridges on the weaker Rove slate, would be the logical course. One could hardly select a more difficult route with more obstacles than the present one.

DULUTH GABBRO (KEWEENAWAN)

The Duluth gabbro lopolith lies between the Rove slate district and the region underlain by the Keweenawan flows and sills. F. F. Grout[3] shows that the Duluth gabbro is composed of an alternation of mineralogically unlike bands, accompanied by a fluxion structure and in some places by sheet jointing. Some of the layers, especially near the bottom, are rich in olivine, which weathers rapidly. The ridges and valleys on the gabbro are the result of differential erosion of the bands composed of more easily weathered minerals and belts of more resistant "red rock," a siliceous differentiation product. The structure of the gabbro turns with the lakes from Brule Lake southwestward to Sawbill Lake. This linear adjustment of the drainage lines to the structure is not, however, so well marked in all parts of the gabbro area, especially not in the area to the southwest of Gunflint Lake.

The Duluth gabbro upland, standing well above the Rove slate lowlands to the north, forms a definite escarpment at the contact with that formation. The topography on the gabbro is that of an undulating plain, below 1,000 feet altitude, for the most part. The region is one of low relief, but steep-sided rock hills and vertical escarpments of 10–20 feet in height are very common and well shown along the Gunflint Trail. The sharp hogbacks of low elevation occur so abundantly as to give a washboard effect. The glacial lakes, none of them very deep, cover a considerable portion of the area.

EFFECTS OF GLACIATION

Glacial erosion has been vigorous; in fact, the entire region was pre-eminently one of glacial scour. The hills are almost without glacial debris, and the valleys contain very little, compared with morainal areas farther south in Minnesota, Wisconsin, and Michigan. The topography is one developed by river sculpture, modified by vigorous glacial erosion and some deposition. Characteristic of the area are the expanses of bare rock scoured by glacial action and the thinness of the soil, which, over large areas, is composed of only a veneer of decayed vegetation, such as a mat of moss and pine needles, or of bouldery, infertile transported material. A striking feature is the almost complete absence of any local or residual soil, such as one would expect on a peneplaned surface of low relief.

The drainage of the region is imperfect. The numerous lakes, swamps, waterfalls, and rapids are features of im-

[3] "A Type of Igneous Differentiation," *Jour. Geol.*, Vol. XXVI (1918), pp. 626–58.

mature drainage. The ice sheets removed the residual soil from the peneplain and brought in a thin, bouldery soil, scoured the ledges bare, and displaced the stream courses from the zones of weaker rocks, producing many waterfalls and rapids. The subsequent longitudinal valleys were clogged by glacial drift, thus forming one type of lakes, while those valleys which were deepened by scour produced another type.

The ice came from almost due north, crossing the ridges at a high angle. Thus parallel valleys were not deepened very much, although in many places glacial scour has made rock basins.

In the Saganaga granite area, numerous rounded, deeply grooved or smoothly scoured, domelike or elongate *roche moutonnée* forms have been developed. The same effect is evident on the monoclinal diabase-sill and dike ridges.

Somewhere in their courses, most of the streams in the region exhibit rapids, waterfalls, and narrow gorges; and doubtless there are streams which are entirely new since the ice age. The length, depth, and width of the postglacial gorges are testimonials to the rapidity of erosion since the recession of the ice sheet. The valley of the Pigeon River below Partridge Falls, with its numerous falls, rapids, and deep narrow gorges, indicates that its present lower course is probably a recent one, a result of its diversion by the ice from a shorter course to the south in preglacial time. It is possible that the preglacial Pigeon River drained through one of the broad, open valleys farther south. The writer, however, has also suggested other possible routes during preglacial time.

ACKNOWLEDGMENTS.—The field work was carried out during the summer of 1946 under a grant from the Robert Wilson Research Fund at the College of Wooster. The writer is indebted to Dr. F. F. Grout and Dr. G. M. Schwartz, of the Minnesota Geological Survey, for information and the loan of manuscripts.

A TENTATIVE CLASSIFICATION OF THE MAIN STRUCTURAL UNITS OF THE ANATOLIAN OROGENIC BELT

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ABSTRACT

A brief summary has been given, in five tables, of the main structural and stratigraphic units of the Anatolian orogenic belt and their possible continuation in southern Europe (the Balkans and northern Italy), based on the author's field observations in the Lombardic Alps (1929–31), Egypt (1937–40), Syria and Palestine (1943), and Turkey (1940–42) and on further evidence from the literature. The main clue to this comparative analysis of regional tectonic units in southeastern Europe and Asia Minor was obtained from a study of the boundary between the so-called "Arabian and Anatolian facies" in southern Turkey and the location of a regional fault zone in northern Turkey, the "Izmit-Erzurum line" (comparable with the San Andreas fault in California), containing the epicenters of devastating earthquakes. Stratigraphical and structural evidence indicate the continuation of this fault zone through the Struma and Morava valleys of the Balkans into the "Drauzug" of the Karawanken Mountains and the Insubric fault zone of the Val Tellina (northern Italy).

INTRODUCTION

There have been several attempts to make a geotectonic analysis of Turkey and neighboring areas, but practically all these studies have been based on scattered geological data published in different articles by geologists who visited Turkey for only a short period. If we realize that even now, after a more or less systematic geological mapping of the country by the *Maden Tetkik ve Arama Enstitüsü* during the last five years, the greater part is still only superficially known, it seems probable that none of these analyses can withstand serious criticism. Turkey is an extremely large country (about 450 km. wide and about 1,500 km. long) with a great variety of most complex stratigraphical and structural problems which can be solved only after years of careful mapping and drilling.

The first structural analysis based on personal observations covering a period of several years was published by P. Arni in 1939.[1] A comparable analysis was published later by Professor W. Salomon-Calvi.[2] Since the publication of these articles, a great number of new data have been collected during exploration work by the Petroleum Department of M.T.A. Enstitüsü, under the leadership of Mr. Cevat Tamşan. Besides surface studies, several deep wells were drilled, which greatly helped to solve certain stratigraphical problems of southeastern Turkey. The writer, who has worked for the Petroleum Department since 1940 both as subsurface geologist and as micropaleontologist, has had the opportunity of visiting many parts of Turkey and of comparing the published data with his own field observations. The following tentative classification of the main structural units of the Anatolian orogenic belt represents a compilation of the data available at present. Undoubtedly, as more information accumulates, changes will be necessary, but our present knowledge makes it rather probable that the main outline will stand. It is for this reason that we have the courage to publish this tentative classification.

[1] "Tektonische Grundzüge Ost-Anatoliens und benachbarter Gebiete," *Maden Tetkik ve Arama Enstitüsü Publication*, Ser. B., No. 4 (1939).

[2] "Kurze Übersicht über den tektonischen Bau Anatoliens," *Maden Tetkik ve Arama Enstitüsü Publication, Bull. 1940*, pp. 49–52.

DIVISIONS AND THEIR CLASSIFICATION

The Anatolian orogenic belt and neighboring areas can be divided into six main units: From north to south, the Pontic orogen, the central Anatolian plateau, the Toros orogen, the Anatolian-Iranian foredeep, the Arabian plateau, and the Arabian shield. Each of these main structural units can be subdivided into smaller units as follows:

I. Pontic orogen or Pontides (P. Arni, ftn. 1)
   A. Pontic foredeep: Kuban depression, Sea of Asow, etc.
   B. Northern Pontic geanticline: central Caucasus, Crimea, etc.
   C. Southern Caucasian trough: Baku-Batum depression, Black Sea, southern Caspian Sea, etc.
   D. Southern Pontic geanticline: northeastern Thrace, Black Sea coastal area, in many places with thrusting toward the north, areas south of the Caspian Sea thrusts mainly toward the north
   E. Northern volcanic zone: Alagöz (4,094 m.), Elbrus (5,631 m.), etc.
   F. Northern, Anatolian fault zone (="Tonale linie" of W. Salomon-Calvi or "Paphlagonische Narbe" of E. Nowack)
      Ismit-Erzurum line
II. Central Anatolian plateau or Anatolides (P. Arni, ftn. 1)
   A. Western and northwestern Turkey (="lydisch-karisches Massif" of W. Salomon-Calvi, ftn. 2)
   B. Konya-Kirsehir Massif (="Galatisch-Lykaonisches Massif" of W. Salomon-Calvi, ftn. 2)
   C. Miocene inner basins of Sivas, etc.
III. Toros orogen or Torides (P. Arni, ftn. 1)
   A. Southern volcanic zone: Hasan Dağ (3,258 m.), Melendizdağ, Erciyasdağ (3,916 m., still active about 100 B.C.), Bingöldağ (3,650 m.), Nemrutdağ (2,949 m., still active in 1441), Ararat (5,165 m.), etc. Volcanic rocks mainly andesites and basalts but also trachytes, etc.
   B. Central Toros geanticline
      1. Antalya Toros (=Pamphylic Toros of Blumenthal)
      2. Seydişehir Toros
         a) Suitandağ
         b) Eğridir-Akseki zone (Seytandağ, Küpedağ, Karakuşdağ, and the southeastern continuation=Imbaros Toros); zone with complicated imbricated structures, all thrust toward the northeast; main serpentine intrusions are Paleozoic
      3. Seyhan Toros
         a) Bulgar-Aladağ zone (=Central Toros or "Kappodokian Toros" of Blumenthal)
         b) Mountains northwest of Seyhan Nehri (=Kilikian Toros); zone with mainly large vertical movements; main serpentine intrusions at the end of the Turonian
      4. Hekimhan-Divrik Toros
         a) Zone with pre-Turonian serpentine intrusions; Middle Eocene syenite and granodiorite intrusions
         b) Ergani Toros: zone with post-Cretaceous serpentine intrusions
      5. Van Toros
         a) Zone with complicated imbricated structures east and south of Lake Van, partly thrust toward the north; many Senonian intrusions
         b) Bitlis-Hakkari Toros (=part of the Iranides of Arni); zone with complicated imbricated structures in the west, small nappe-overthrusts in the east, and all directed toward the south; main serpentine intrusions at the end of the Turonian
   C. Antalya-Adana-Diyarbakir troughs
   D. Southern Toros geanticline
      1. Cyprus (=Trodos complex)—Hatay Toroş (Kizildağ and Gavurdağ="Amanos" of previous authors)—Maraş (=Pyramian Toros) zone; large-scale serpentine intrusions in pre-Turonian time
      2. Siirt-Mardin-Beytuşşebap zone; anticlinal block structures, partly thrust toward the south, separated by large longitudinal fault zones
      3. Zagros range (in Iraq). The Kilikian Toros and Pyramian Toros together are also called "Anti-Toros"

E. Boundary between Arabian and Anatolian facies, often accompanied by a steep folded zone or by longitudinal faults with large-scale basalt extrusions.

IV. Anatolian-Iranian foredeep
  A. Mesopotamian foredeep: Musul-Cizre-Nusaybin-Urfa-Gaziantep zone; near Musul most anticlines asymmetric with steep flank on southwestern side
  B. Gebel Sincar geanticline: Gebel Sincar-Abdul Aziz zone

V. Arabian plateau: Region south of the Gebel Sincar geanticline as far as the Egma Plateau in southern Sinai; many block folded structures and regional graben zones in an over 2,000-m.-thick Mesozoic-Tertiary section

VI. Arabian shield

The writer had intended to indicate these different zones on a map showing their relation to the structural lines of Greece, Bulgaria, Syria, Iraq, and Iran. But as so many problems are still unsolved and particularly as the boundaries between the different units are known accurately in only a few cases, he finally thought it better not to present such a map. Too many maps have been published already on the geotectonics of this part of the world which have been copied in textbooks without further consideration.

A detailed description of each of the above-mentioned units is beyond the scope of this article, especially as a tectonic analysis of each zone should be preceded by a stratigraphical analysis. To provide some of the necessary information, we have prepared five stratigraphic tables (Pls. I–V) giving the main lithologic characteristics of those areas which have been recently studied in detail. These tables also indicate which areas in Turkey are rather well known and which ones require more detailed studies. The vertical distances in the stratigraphic tables were made in accordance with the absolute time scale. From the tables the differences in rate of sedimentation in the different periods can easily be deduced.

The classification of the different tectonic units and particularly the boundaries between them are based mainly on two fundamental structural lines in Turkey which have been studied in detail during recent years—the so-called "Izmit-Erzurum line" and "the boundary between the Anatolian and Arabian facies." As the exact position of these lines is of fundamental importance in understanding the geotectonic structure of Turkey, it is necessary to discuss each of them in detail.

THE BOUNDARY BETWEEN ARABIAN AND ANATOLIAN FACIES

In the Upper Cretaceous and Eocene sections of southern Turkey two facies can be distinguished: a northern facies composed of gray or dark-gray marls, which the writer has called the "Anatolian facies"; and a southern facies, composed of glaring white chalks, chalky marls, chalky limestones, etc., to which we have given the name of "Arabian facies" because a similar facies is known from Syria, Palestine, and Egypt. The boundary between these was accurately established by field studies in the Urfa and Gaziantep region by I. Ortynski and V. Stchepinsky. This facies boundary and the microfaunal characteristics of the Arabian facies have been described recently in an article by Tromp.[3]

The boundary between Arabian and Anatolian facies is mostly located along a steeply folded zone or by longitudinal faults, accompanied in many places by extensive basalt extrusions. This indicates that the boundary is due to a tec-

[3] "Micro-faunas of the Upper Cretaceous and Tertiary Sections (Arabian Facies) in the Urfa and Gaziantep Regions," *Maden Tetkik ve Arama Enstitüsü Publication, Bull. No. 8* (1943), pp. 134–41.

tonic cause. As explained in the above-mentioned article, a notable difference in open-sea conditions, which involved more than a difference in depth of the sea, must have developed during Upper Cretaceous and Eocene times between southern Turkey and the areas farther north.

The well-marked boundary runs in southeastern Turkey almost east-west along the steep mountain front, north and northeast of Cizre, along Mardin southwest of Derik, whence it swings toward the northwest, runs east of the line Viranşehir-Siverek, swings north of Siverek to the west, and goes from there north of Adiyaman toward the southwest, just west of Burç (18 km. west of Gaziantep) and west of Kilis into Syria.

The areas south of this boundary roughly coincide with the northern part of the so-called "Arabian shield," as we pointed out in the above-mentioned article. This name, however, is not strictly correct, as in northern Egypt, Palestine, and Syria a Mesozoic and Tertiary section of over 2,000 m. was deposited on old Paleozoic rocks and was steeply folded into dome-shaped structures. The whole area has, therefore, more the characteristics of a geosynclinal region than of a shield; but because, on the other hand, these areas reacted rather rigidly against tectonic forces (compared with Turkey), we proposed the name "Arabian plateau" for this region, which, still farther south (from the southern part of the Sinai peninsula toward Arabia), passes into the real "Arabian shield" of dominantly pre-Cambrian rocks.

However, the areas south of the facies boundary, in southern Turkey, do not all belong to the Arabian plateau. Immediately south lies a deep basin with a thick sedimentary section, which, except for the Upper Cretaceous and Eocene, does not differ fundamentally from that farther north. It is a zone with many anticlinal structures, particularly in Iraq and vicinity, where they are rich in oil seepages. The whole oil-field belt of Iraq, with the Mesopotamian depression, belongs to this zone. It continues through the Persian Gulf, south of Iran, into the Indus and Ganges depression south of the Himalayan mountains. From there it runs west of Burma, probably along the Andaman and Nicobar islands, toward the islands west of Sumatra. Farther eastward it follows the zone with earthquake epicenters and large negative anomalies south of Java. These large negative anomalies all along this zone in the Indian Ocean (discovered by F. A. Vening Meinesz), their continuation in the so-called "Plumb line" in India (also with large negative anomalies), and the presence of many earthquake centers indicate that great parts of this zone are still in a stage of upfolding. As this zone, which is generally accepted as the southern foredeep of the Himalayan and Malayan orogenic belt, continues into the Mesopotamian depression, we have called the portion of this belt immediately south of the Anatolian-Arabian facies boundary "the Mesopotamian foredeep." The southern boundary of this foredeep in Syria lies probably at the northern flank of the geanticlinal zone formed by Gebel Abdul Aziz and Gebel Sincar. This geanticline swings toward the southeast, following the west border of the Tigris depression. It is the area south and west of this geanticlinal zone which we have called the real Arabian plateau.

THE IZMIT-ERZURUM LINE

Location of the Fault Zone in Turkey

In northern Anatolia an enormous fault zone occurs, similar to the San Andreas fault zone in California, which di-

AGE (vertical distances in accordance with absolute age in million years)				ANATOLIAN-IRANIAN FORE DEEP: 1) Urfa-Gaziantep region		SOUTHERN TOROS GEANTICLINE: 2) SE Turkey (N of Syrian plain)		3) Hatay (& N W Syria)	
				Lithology	Thickness in meters	Lithology	Thickness in meters	Lithology	Thickness in meters
CENOZOIC	QUAT.	PLEISTOCENE (8 MY)		Basalt flows	5-10	Basalt flows	5-10	Basalt flows, thick congl	?
CENOZOIC	TERTIARY (58 MY)	Neogene	Pliocene	Reddish and greenish marls alternating with white porous limestones. Fossils rich in gastropods etc	< 100	Grey, loosely cem. S.S., in lower part calcareous with marly interc., several congl interlayers	> 500	1) Alternation of s.s., marls and congl (with serp. pb.) 2) l.g marls	> 350
CENOZOIC	TERTIARY (58 MY)	Neogene	Miocene U	Creamish, reddish and yellow limestones. Fossils rich in oysters, Pecten, Lepidocyclines etc	< 100	Whi. L.S. near base alterns. of g. gr. & l.r. marls and S.S.	> 130	l.gr.g marls with thin s.s. interc	> 400 ?
CENOZOIC	TERTIARY (58 MY)	Neogene	Miocene M			g. gr. & l.r. gyps marls with L.S. interc.	> 500	l.g. marls, near base often shading into L.S.	> 300
CENOZOIC	TERTIARY (58 MY)	Neogene	Miocene L			Whi. gypsum with shale int	> 250	Missing	—
CENOZOIC	TERTIARY (58 MY)	Palaeogene	Oligocene	White, creamish and brownish limestones. Fossils first lepidocyclines last camerina	< 50	Missing	—	Probably missing. Diabasic extrusions, age ?	—
CENOZOIC	TERTIARY (58 MY)	Palaeogene	Eocene U	Cr. whi. and whi. chalk, chalky marly L.S. and limestone	> 100	Missing	—	Missing	—
CENOZOIC	TERTIARY (58 MY)	Palaeogene	Eocene M	Whi. dense L.S. rich in chert. Fossils camerina, orbitoclyclin	200	Middle L.S. whi. massive L.S. rich in chert	250	l.g. L.S. rich in chert, camerina etc. in upper part chalky	?
CENOZOIC	TERTIARY (58 MY)	Palaeogene	Eocene L	Whi. chalky marly L.S. Fossils Marly Echinoids	350	Lower Red Beds & Flysch series, mainly of gray sandy marls	> 400 450	Missing	—
MESOZOIC	CRETACEOUS (65 MY)	Upper Cretaceous	Senonian	l.g & cr whi. chalky marly L.S. Alternation of white & cr whi chalky marly L.S and chalky limestone	235	g. or d.g marls, near base few thin L.S. interc with orbitoides & amphalocyclus	abt. 200	Practically missing at base orbitoidal marly L.S	?
MESOZOIC	CRETACEOUS (65 MY)	Upper Cretaceous	Turonian	Grey massive limestone near top Hippurites	?	Top Classy beds g marly L.S. altern with thin b.d limestones. Lower section: g & massive L.S near top chert and rich in Hippurites	> 30 abt 200	Top marls & thin bd marly L.S. Lower part g Hippuritic limestones	?
MESOZOIC	CRETACEOUS (65 MY)	Upper Cretaceous	Cenomanian (+Albian & Aptian)	Cenomanian and older beds not exposed but probably present	?	Dark grey & g dolomite with limestone interc	abt 200	1) limestones with marl int 2) ye & whi oblitinc L.S with Orbitolina 3) gr nodular marly L.S	> 400
MESOZOIC	CRETACEOUS (65 MY)	Lower Cretaceous		?	?	d.g & g dolomite with limestone intercalations	abt 200	Probably missing	—
MESOZOIC	JURASSIC (32 MY)	Malm		?	?	Dark grey and black limestones with dolomitic intercalations; several limonitic zones	> 400	1) L.S. partly nodular, ferruginous & rich in lamellibr 2) L.S. with marl interc near the base (only exposed in Syria)	?
MESOZOIC	JURASSIC (32 MY)	Dogger		?	?			1) marly L.S 2) algae L.S (sectionally exposed in Syria)	> 1080 (?)
MESOZOIC	JURASSIC (32 MY)	Liassic		?	?			Probably missing	—
MESOZOIC	TRIASSIC (28 MY)	Upper Triassic		?	?	Goyan series. Thin bd g & gr g argillaceous L.S., marly L.S with many g claystone interc., in lower part many r & gr. arg L.S (shaly weath) interc. Fossils. Pseudomonotis clara	170 - 300	Probably missing	—
MESOZOIC	TRIASSIC (28 MY)	Middle Triassic		?	?				
MESOZOIC	TRIASSIC (28 MY)	Lower Triassic		?	?	g. d.g extremely massive L.S	> 150		

Column

1. Based on field studies of I. Ortynski (1940–43).
2. Based on surface and subsurface data studied by S. W. Tromp (1940–43); field studies by C. Tamsan, Kemal Lokman, Vonderschmidt, Kirk, Maxson, Arni, Cunningham Craig, Ericson, Foley, Moses, Paige, Schmid, Sparks, McCammon, Miss Jordan, Ortynski (1935–43).
3. Based on surface data of Ericson, Wijckerslooth; surface and subsurface studies by S. W. Tromp (1940–42).
4. Based on field studies by E. Lahn, I. Ortynski, and S. W. Tromp (1941).
5. Based on field studies by M. Blumenthal, E. Foley, and S. W. Tromp (1941–43).
6. Based on field studies by P. J. C. de Wijckerslooth (1941).
7. Based on field studies by J. H. Maxson (1937), I. Ortynski, and S. W. Tromp (1943).
8. Based on field studies by P. Arni (1938), E. Lahn (1939), E. Schaput (1936), M. Blumenthal (1942), and S. W. Tromp (1942).
9. Based on field studies by M. Blumenthal (1938).
10. Based on field studies by I. Ortynski and S. W. Tromp (1942).
11. Based on field studies by P. Arni (1930), G. Ralli (1933), and W. Grancy (1938).
12. Based on field studies by Hamit Pamir, E. Paréjaz, L. von Loczy, and W. Paeckelmann, and subsurface studies by S. W. Tromp (1943).

The following symbols apply to all the tables: ~~~~ = unconformities; < = less than; > = more than; *arkose*, feldspar sandstone; *graywacke*, gray, brown, and black conglomeratic arkose with angular shale pieces, etc.; *cr.*, creamish; *whi.*, white; *gr.*, green; *r.*, red; *b.*, black; *bl.*, blue; *br.*, brown; *g.*, gray; *l.*, light; *d.*, dark; *L.S.*, limestone; *S.S.*, sandstone; *cem.*, cemented; *congl.*, conglomeratic; *interc.*, intercalation; *intr.*, intrusions; *serp.*, serpentine; *pb.*, pebbles; *gyps.*, gypsiferous.

NOTE: In Col. 1 of Table I, below the Middle Eocene unconformity, are about 100 m. of whitish chalky limestones, which represent the basal Middle Eocene. This was left out for lack of space. In Col. 2 of Table I, below the Middle Eocene Midyat limestone, two Middle Eocene units occur which could not be described for lack of space: the Gercüs red beds (up to 270 m.), composed of red clays, gypsiferous shales, red sandstones, and conglomerates; they are underlain by the flysch series (more than 150 m.), composed of gray sandy marls alternating with greenish-gray calcareous sandstones. The base of the flysch series mostly coincides with the base of the Middle Eocene, but the upper part of the Lower Eocene is sandy in places.

II

AGE (vertical distances in accordance with absolute age in million years)				CENTRAL TOROS GEANTICLINE							
				4) Seydişehir - Toros		5) Seyhan Toros		6) Divrik Toros		7) Bitlis-Hakkari Toros	
				Lithology	Thickness in meters	Lithology	Thickness in meters	Lithology	Thickness in meters	Lithology	Thickness in meters
CENOZOIC	QUAT.	PLEISTOCENE (2 M.Y.)		Volcanic tuffs & freshwater lake dep		Hard cem congl (igneous pb)	> 600	?	?	congl deposits & Basalt flows.	
CENOZOIC	TERTIARY (58 M.Y.)	Neogene	Pliocene	Thick tuff deposits, Andesite & trachyte extrusions	?	Missing		?	?	S s & congl (composed of Permocarb. Cret etc) In upper part andesite & tuff int	> 1000
CENOZOIC	TERTIARY (58 M.Y.)	Neogene	Miocene U	Altern of g marls & L s serp congl at the base	> 100	(N of Adana) Altern of g marls & s s	600 - 800	?	?	only near S. border: 1) whi L s 2) g.grg & r marls & s s	> 450
CENOZOIC	TERTIARY (58 M.Y.)	Neogene	Miocene M	Missing	–	g marls with s s interc	> 100				
CENOZOIC	TERTIARY (58 M.Y.)	Neogene	Miocene L			calc s s with marl interc	1000 - 1200			g.grg & r gyps marls	> 500
CENOZOIC	TERTIARY (58 M.Y.)	Neogene	Miocene L			1) g marls with thin s s int 2) g & yc br s s & congl	2600 1300 - 2600	Basalcongl with Syenite bld		Missing	–
CENOZOIC	TERTIARY (58 M.Y.)	Palaeogene	Oligocene	Missing	–	g limestone	100 - 200	Probably missing	–	Missing	–
CENOZOIC	TERTIARY (58 M.Y.)	Palaeogene	Oligocene	Missing	–	Ecemis Cay series 1) g & gr g sandy marls with s s int 2) s s and congl (Oligocene missing N of Adana)	> 200				
CENOZOIC	TERTIARY (58 M.Y.)	Palaeogene	Eocene U	Missing	–	Missing	–	Syenitic & granodiorite intr formation of magnetite dep	?	Missing	–
CENOZOIC	TERTIARY (58 M.Y.)	Palaeogene	Eocene M	l.g-whi Camerina L s	?	1) camerina limestone 2) calc s s and congl	?	S s & thin bd sandy L s with Camerina etc	?	whi & b limestone rich in Camerina & fasciolites etc brecciated L	> 100
CENOZOIC	TERTIARY (58 M.Y.)	Palaeogene	Eocene L	Probably missing	–	Probably missing	–	Missing	–	M.E. 1) Gercüs Red Beds r S s & congl 2) Flysch series Thick bd gr s s & s marls intr by serp	
MESOZOIC	CRETACEOUS (65 M.Y.)	Upper Cretaceous	Senonian	Practically missing at base. g L s with orbitoids	–	Missing or extremely thin		Missing	–	L Eocene missing (Miocene directly on serp congl) Senonian sed rare Red bld conglomerate Large serpentine intr.	?
MESOZOIC	CRETACEOUS (65 M.Y.)	Upper Cretaceous	Turonian	l.g limestones with Tenagodes, Hippurites etc	> 100	? ? ? ? large gabbroidic intr 1) r & gr thin bd L s with marl int & r chert nodules 2) Thick bd l.g L s with Tenagodes	? > 100	g Rudist-limestone	?	d.g unfossiliferous L s	?
MESOZOIC	CRETACEOUS (65 M.Y.)	Upper Cretaceous	Cenomanian (+Albian & Aptian)	(only developed near Akseki) l.br. argillaceous L s with Gryphea and Exogyra spec	?	whi or g massive dolomitic limestone	> 100	g to b massive dolomitic L s	?	d.g massive dolomitic L s (?)	?
MESOZOIC	CRETACEOUS (65 M.Y.)	Lower Cretaceous		Probably missing small scale serpentine intr	–	Missing		Missing Large scale basic intr	–	Probably missing	–
MESOZOIC	JURASSIC (32 M.Y.)	Malm		1) Red series (+ Radiolarite series) Red shales with many r & gr chert 2) Flysch series altern of g sandy (int shales & marls & micac s s	> 200 > 1000 max	Radiolarite series r & whi L s with r chert altern with gr marls & calc s s	? few hundred	Thin bd L s with radiolarite(?) intercalations Age unknown	?	Missing	–
MESOZOIC	JURASSIC (32 M.Y.)	Dogger		g massive L s	> 200	l.g limestone (?)	?	?	?		
MESOZOIC	JURASSIC (32 M.Y.)	Liassic		Flaggy limestone series b thin bd L s with b shaly marl intercalations	> 250	Probably missing	–				
MESOZOIC	TRIASSIC (28 M.Y.)	Upper Triassic		Phyllitic series: micaceous phyllites with many quart int (strongly squeezed in écailles)	> 400	Megalodon limestone d.g thin bd L s with megalodon prints (Norian ?)	? > 1000	?	?		
MESOZOIC	TRIASSIC (28 M.Y.)	Middle Triassic		Probably missing	–	?	?	?	?	Missing	–
MESOZOIC	TRIASSIC (28 M.Y.)	Lower Triassic		Probably missing	–	?	?	?	?		

AGE (vertical distances in accordance with absolute age in million years)				8) CENTRAL ANATOLIAN PLATEAU (between Konya Çankırı & Kırşehir)		5 SOUTHERN PONTIC GEANTICLINE							
						9) Ilgaz Mts		10) Boyabat - Sinop region		11) Zonguldak region (coastal area between Zonguldak & Ereğli)		12) Trakia	
				Lithology	Thickness in meters	Lithology	Thickness in meters	Lithology	Thickness in meters	Lithology	Thickness in meters	Lithology	Thickness in meters
CENOZOIC	QUAT.	PLEISTOCENE (2 M.Y.)		Volcanic tuffs & freshwater lake deposits			—	Basalt flows	> 10			Basalt flows	?
CENOZOIC	TERTIARY (58 M.Y.)	Neogene	Pliocene	Thick tuff deposits Andesite & basalt extr		Missing	—	Missing	—	Missing	—	r.ye & br soft ss with clay intercalations (Central Thrace) or congl & ss with Dreissensia etc (S W Thrace)	?
CENOZOIC	TERTIARY (58 M.Y.)	Neogene	Miocene U	Marls with gypsum, with lignite (Sinop)	to E Turkey marine deposits (Ispidergel)	?						S W Thrace: Fresh water & water Ls, shale, gypsum, lignite	?
CENOZOIC	TERTIARY (58 M.Y.)	Neogene	Miocene M	r.ligni ss and marls, gypsum with lignite		> 1000						W Thrace: r.ye, red soft s.sand, marl, lignite, gypsum, white marls all with gypsum	
CENOZOIC	TERTIARY (58 M.Y.)	Neogene	Miocene L	Probably missing	& Miocene (Ispidergel)	—						Probably missing	—
CENOZOIC	TERTIARY (58 M.Y.)	Palaeogene	Oligocene	Andesitic series (now unknown) (Kırşehir) (Sapolir) ss, at the base thick rockfall deposits Andesitic series, tuffs, clay, ss with many lignitic beds	> 400 > 50	Missing (present in Kastamonu - Boyabat depression further N)	—	ss-s m thick gr clay bd with s m thick congl int (pb of serp & Eocene Ls	> 200	Missing	—	1) Sandstone and shale with coal seams and few volcanic tuff interc 2) g Ls with g thin bd partly cherty, shale int, few congl	> 1000 ?
CENOZOIC	TERTIARY (58 M.Y.)	Palaeogene	Eocene U	Massive ss with g & violet marl interca. mostly Discocyclines	> 200	Missing (present in Kastamonu-Boyabat)	—	Discocyclina Ls & lack of g marls & br. tuffs	> 25 400	Missing	—	1) oligin with g marls, marly Ls and few ss indiff	> 180
CENOZOIC	TERTIARY (58 M.Y.)	Palaeogene	Eocene M	Alternation of the part with marl, red	400	Camerina Ls	?	Nummulitic Ls & sandy marls	> 1000	Missing	—	gl/lim bed congl with medium etc	> 200
CENOZOIC	TERTIARY (58 M.Y.)	Palaeogene	Eocene L	Red Ls (Camerina Ls)	> 50	Missing	—	Lutetian Ls with marls at base	> 180	Missing	—	g cut Camerina Ls partly Ls and g marls, finally sandy, medium clay	> 180
MESOZOIC	CRETACEOUS (65 M.Y.)	Upper Cretaceous	Eocene	Probably missing		Practically missing		Practically missing	—	marls, ss & gc marly Ls with interc of andesites	few hundred	Mostly missing Ls & marl with basalt and andesite intercalations	?
MESOZOIC	CRETACEOUS (65 M.Y.)	Upper Cretaceous	Senonian	Practically missing Mar Rudister: b.g.marls and ss with hippuritic intercal	> 50	Sandy Ls, marls and calc ss (Fossils: Inoceramus Orbitoides)	? thin	calc ss (if covering basement) or dg marls with Ls intercal (Inoceramus etc)	< 50	Fossils Inoceramus etc		Fossils Inoceramus etc	
MESOZOIC	CRETACEOUS (65 M.Y.)	Upper Cretaceous	Turonian	1) Sandy Ls with orbitoids (age unknown) 2) white massive Ls with Rudists (E of Ishilip)	?	g Hippurite limestone Fossils Rudists, Actaeonella	?	g white hippurite Ls (massive)	> 50	congl of large Lorch bld, Lst Cret pb	few hundred	Mostly missing Hippurite limestone conglomerate at base (overlying Devonian or Trassic)	?
MESOZOIC	CRETACEOUS (65 M.Y.)	Upper Cretaceous	Cenomanian (+Albian & Aptian)	Ss, marls & r congl with interc of andesites, porfyrites etc (few Cenom fossils) (only developed in the west R part)	?	g limestone Fossils Caprina	?	g l b shales with thin ferruginous ss intercalations Fossils Duvalia, Neohibolites, Inoceramus etc	> 250	1) pattern of marl, marly Ls, ss (with few coal beds) 2) g & g marls with Hippurites etc 3) Glauconite ss rich in fossils 4) Marl with small Rudists 5) b & gr marls, gr clay, r ss, congl & ss (rich in Orbitolines)	few hundred 100 - 200 100 320 150 70	Missing	—
MESOZOIC	CRETACEOUS (65 M.Y.)	Lower Cretaceous		Practically missing locally wh & g Ls with marl intercalations Fossils Duvalia, neocomites serpentine intrusions	—	b-g Ls and b marls (Fossils Desmoceras, Crioceras spec) Prob serpentine intr	?	1) pattern of g marl and br calc ss (large "ripple marks) 2) g shale with 10 cm thick ss interc (ripple marks) 3) b, gr-g marl with ss int 4) b shale with Ls int	350 180 > 200 > 50	1) g Ls rich in Requienia	abt 100	Missing	—
MESOZOIC	JURASSIC (32 M.Y.)	Malm		Thin bd Ls with thin g radiolarite intercal Marly Belemnite fragm	?	Probably missing		Missing	—	(Missing near Zonguldak) 1) Thin & thick bd Ls & oolith 2) A gr ss congl & sandy clay after directly on Permian	600 20 - 200	Missing	—
MESOZOIC	JURASSIC (32 M.Y.)	Dogger		1) Ls with ammonites 2) Marls with Posidonomya	?	g Ls with b hard shale interc, few ammonites	?	Missing	—	(Missing near Zonguldak) 1) m & g br gc sandy clay, r ss with traces of lignite 2) b clay, g gr s rel marls, ss	200 - 300 100 - 150	Missing	—
MESOZOIC	JURASSIC (32 M.Y.)	Liassic		Red Ls rich in ammonites & r shaly Ss, at base congl	?	Probably missing	—	Missing	—	(Missing near Zonguldak) 1) s s & sandy clay (with lignite) 2) gr, r s rel clay & marl	300	Missing	—
MESOZOIC	TRIASSIC (28 M.Y.)	Upper Triassic		Strongly recryst. (g & massive Ls, rich in gastropods (prob Ladinian)	? (thin)	Missing	—	Missing	—	Missing	—	Missing, only developed E of Bosphorus Rhetian - Norian: missing Carnian Ladinian Anisian } massive Ls Werfenian: { sandy marl ss etc congl	?
MESOZOIC	TRIASSIC (28 M.Y.)	Middle Triassic		Missing	—								
MESOZOIC	TRIASSIC (28 M.Y.)	Lower Triassic											

AGE (vert. distances in accordance with absolute age in million years)		ANATOLIAN-IRANIAN FORE-DEEP		SOUTHERN TOROS GEANTICLINE				C E N T
		1) Urfa-Gaziantep region		2) S.E. Turkey (N. of Syrian plain)		3) Hatay (& N.W. Syria)		4) Seydişehir - Toros
		Lithology	Thickness in meters	Lithology	Thickness in meters	Lithology	Thickness in meters	Lithology
PALAEOZOIC	PERMIAN (38 M.Y.)	?	?	Giridağ series: r. br. quartzites gr. argillites b. limestones (Age: uncertain, pre-Goyan series, probably post carb.)	> 450 75 > 30	Probably main period of serpentine intrusion (many chromium dep.) Acc. to Dubertret a.o. age of intrusion U. Cret.	—	1) Reddish and greenish s.s. and congl., composed of serpentine pb. quartz etc. 2) Large serpentine massive
	CARBONIFEROUS (86 M.Y.) Upper Carb. / Lower Carb.	?	?	Harbol series: exact age unknown Black limestones with Productidae and Fusulinidae In lower part alt. with b. shales and ye. br. quartzites. ?	> 250 ?	Missing	—	dg. or b. fusulina and productidae L.s. ?
	DEVONIAN (45 M.Y.)	?	?	Probably missing	—	Missing	—	Black coral L.s
	SILURIAN (94 M.Y.)	?	?	Probably missing	—	? Gavurdağ series: Phyllites, marbles & quartzites (intruded later by serpentines) Fossil: Trilobites	? ?	? ?
	CAMBRIAN (105.000.000 YEARS)	?	?	Telbism series: (at Derik, W of Mardin) 1) g. thick bd. s.s. 2) g. s.s. and green micaceous shales with Brachiopods and Trilobites 3) d.g. thick bd. dolom. L.s. with many chert concretions 4) Thick bd. r. sandst., in lower part thin bd. and gr. s.s. interc. 5) conglomerate of r. & gr. porphyric p.b. Volcanic series: red porphyry flows with few "lahar" & lapilli intercalations (Age: uncertain, perhaps Algonkian Proterozoic)	> 700 40 200 480 40 > 800	?	?	?

AL TOROS GEANTICLINE

5) Seyhan - Toros		6) Divrik - Toros		7) Bitlis-Hakkari-Toros	
Lithology	Thickness in meters	Lithology	Thickness in meters	Lithology	Thickness in meters
Dark grey Fusulinidae limestone	?	?	?	*Practically missing* *d.g. or b. fusulina L.s.*	?
Black limestones with few b. shale intercalations. Fossils: Fusulinidae; rich in Brachiopods (Productidae, Spiriferidae) ?	several hundred	?	?	*1) Black massive L.s. or marbles, often rich in fusulinidae* *2) Alternation of r. & gr. thin bd. L.s. with r. shaly intercal.* ?	>100 50 ?
1) Alternation of br. & b. Spirifer limestone *Fossils: Brachiopods, Corals* *2) Br. calcareous s.s. and sandy L.s.*	probably few hundred	?	?	*Probably missing*	—
?	?	?	?	?	?
Niğde series: *whi. marbles, many quartzites, phyllites, micaschists, gneiss, amphibolites etc.* *(Age: uncertain, but pre-Devonian)*	?	?	?	*Pervari series:* *micaschists, quartzites, (often ferruginous) amphibolite intrusions* *(Age: uncertain, but pre-Carboniferous)*	?
?	?	?	?	?	?

	Thickness in meters
	—
	—
several hundred	
50	
max. 100	
>500	
100-200	
100-200	
70-150	
100	
300-400	
30-40	
100	
50	
?	
	?
	?

vides the mountainous part of Turkey into two large units. This fault zone was called by E. Nowack the "Paphlagonic scar" and the "Tonal line" by W. Salomon-Calvi. The latter, in particular, has written several articles on this fault zone,[4] which he was inclined to consider equivalent to the regional fault zone south of the Alps, called the "Insubric fault zone" by A. Spitz and previously called "Tonal line" by Solomon-Calvi (in 1890). This Insubrian fault zone, considered by Staub and others as the Alpino-Dinaric boundary, was studied in detail by Cornelius and his wife and likewise by the present writer over a considerable distance in northern Italy.[5]

Before considering whether the "Paphlagonic scar" can be the same as the "Insubric fault zone" in Italy, we shall first discuss the characteristics of the northern Anatolian fault zone. This zone comprises a series of parallel thrust faults, which show up topographically in the form of different large depressions ("Ovas"). But most striking is the presence of many earthquake centers all along this zone, which indicate that intermittent movement is still continuing along these faults. The location of these centers of devastating earthquakes, combined with the surface studies of Blumenthal and others, makes it possible to trace this fault zone very accurately. It can be traced with certainty (Fig. 1) through the Gulf of Izmit, along Izmit, Sapança Gölü, south of Adapazari and Hendek, south of Bolu, along the Gerede Çay (south of Arkot Dağ), along the Upper Ulu Su and Devrez Çay, south of the Ilgaz Mountains, along the valley south of Tosya, along the Kizil Irmak depression, via Osmancik, and along the Sulu Ova to Erbaa; thence along the Kelkit Irmak to Niksar, Reşadiye, south of Koyulhisar, north of Suşehri, and north of Refahiye and Erzincan. A cross-fault brings the fault line farther north toward Tercan and Erzurum. From there it continues in a northeast-southwest direction via Hasan Kale (Pasinler), Kağizman to the Aras Graben Valley (north of the Ararat volcano and south of the Alagöz volcano), and along this valley in a southeast direction through Russia into western Iran. As this fault zone lies between the well-known cities of Izmit and Erzurum, it may be designated appropriately as the "Izmit-Erzurum line."

CONTINUATION OF THE FAULT ZONE WEST OF TURKEY

The continuation of this fault zone west of Izmit is not completely certain, but most probably it follows the more than 1,000-m. deep east-west troughs in the Sea of Marmara and crosses the Gallipoli Peninsula into the Gulf of Saros. From there it probably follows the southwest-directed depressions in the sea between Samothrace and Imbros Island. Farther westward it swings to the northwest but is probably cut by a regional northwest-southeast fault zone. This western continuation into Greece is uncertain, and we shall come back to it later after the discussion of the Toros orogen. It is most likely that it runs east of the Vardar Valley in Greece, probably near or in the Struma Valley, and from there in a northwest direction

[4] "Fortsetzung der Tonalelinie in Klein Asien," *Akad. d. Wiss. Wien*, Vol. XVII (June, 1937); ftn. 2; "Untersuchungen über Erdbeben in der Türkei," *Maden Tetkik ve Arama Enstitüsü Publication*, Ser. B, No. 5 (1940).

[5] H. P. Cornelius and M. Furlani-Cornelius, *Die insubrische Linie vom Tessin zum Tonale Pass* (Wien, 1930); S. W. Tromp, *La Géologie du valle del Bitto et la tectonique des Alpes lombardes* (Leyden, 1932); "De Geologie der Bergamasker Alpen in Verband met het Dinaridenprobleem," *Geologie en Mynbouw*, November 16, 1932; "De Genese der lombardische Alpen en die der oost alpine Dekken," *ibid.*, December 16, 1932.

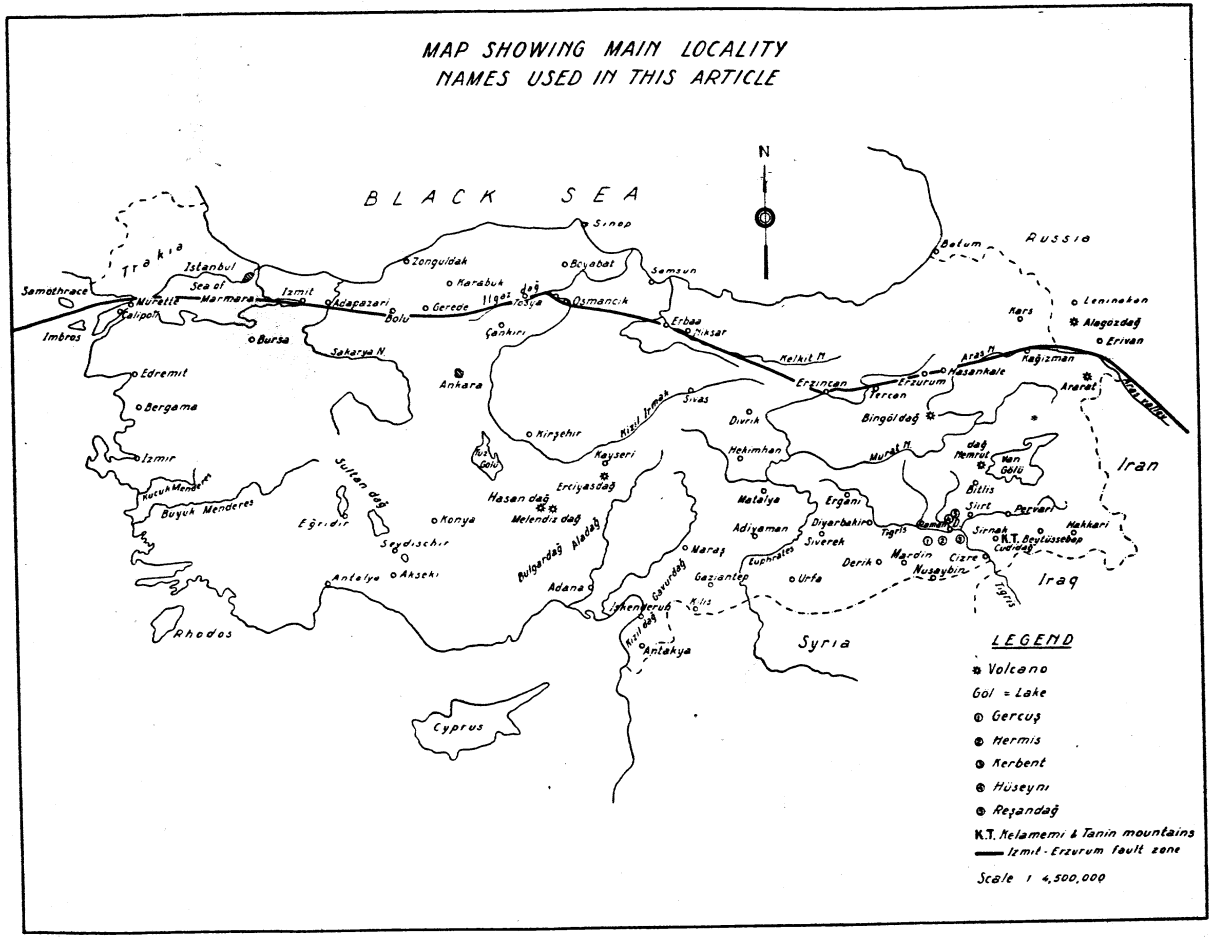
MAP SHOWING MAIN LOCALITY
NAMES USED IN THIS ARTICLE
N
BLACK SEA
Russia
Iran
Iraq
Syria
Cyprus
Rhodos
Trakia
Samothrace
Imbros
Istanbul
Sea of Marmara
Murette
Calipoli
Izmit
Adapazari
Bolu
Gerede
Zonguldak
Karabuk
Ilgaz dağ
Tosya
Osmancik
Çankiri
Sinop
Boyabat
Samsun
Erbaa
Niksar
Batum
Kars
Leninakan
Alagozdağ
Erivan
Kağizman
Ararat
Kras valley
Aras N.
Hasankale
Erzurum
Tercan
Erzincan
Kelkit N.
Bingöldağ
Bursa
Sakarya N.
Ankara
Edremit
Bergama
Izmir
Kucuk Menderes
Buyuk Menderes
Sultan dağ
Eğridir
Konya
Seydischir
Antalya
Akseki
Kirşehir
Kizil Irmak
Sivas
Divrik
Hekimhan
Kayseri
Erciyasdağ
Hasan dağ
Melendiz dağ
Bulgardağ
Aladağ
Adana
Maraş
Gaziantep
Gavurdağ
Kilis
Antakya
Kizil dağ
Iskenderun
Urfa
Malatya
Adiyaman
Siverek
Diyarbakir
Ergani
Tigris
Euphrates
Derik
Mardin
Nusaybin
Cizre
Ramani
Siirt
Bitlis
Van Gölü
Nemrut dağ
Pervari
Sirnak
K.T. Beytüssebap
Cudidağ
Hakkari
Tigris
Kuz Gölü
Kelamemi & Tanin mountains
Izmit - Erzurum fault zone
Scale 1 4,500,000
LEGEND
Volcano
Göl = Lake
Gercuş
Hermis
Kerbent
Hüseyni
Reşandağ

FIG. 1

through the Morava Valley in Yugoslavia.

From the main Izmit-Erzincan fault zone, other fault zones branch off in different directions, but they cannot be discussed for lack of space.

AGE OF THE FAULT ZONE

The age of the main fault is probably pre-Cretaceous, as it disappears in places under Cretaceous flysch. The fault zone may have been present even in Paleozoic time. The Insubric fault zone in Italy is along a Paleozoic fault line, on which there was renewed movement during the Tertiary folding.

SUMMARY OF DEVASTATING EARTHQUAKES IN TURKEY

As stated previously, one of the most characteristic features of this fault zone is the occurrence of many destructive earthquakes all along the zone. Between 1939 and 1943 four severe earthquakes took place: at Erzincan (21.XI.1939 at 10 o'clock and 27.XII.1939 at about 2 o'clock); near Osmancik, about 120 km. south of Sinop (at Kargi on 21.XI. 1942, in the Kizil Irmak Valley on 2.XII. 1942, and in the Hamamözu Valley on 11.XII.1942); near Erbaa and Niksar, about 70 km. south of Samsun (20.XII. 1942), and near Adapazari (20.VI.1943 at 18 hr. 33 min. and 19 hr. 48 min.). Hundreds of people were killed during these earthquakes, and thousands of houses were destroyed.

The Erzincan earthquake was followed a few days later by recurring shocks at several places along the Izmit-Erzurum line as far as Mürefte. This is another indication that these different epicenters really belong to the same fault zone. Some portions of this fault zone, which is over 1,500 km. in length, are more active than others. Izmit, for example, has several times been partly or completely destroyed (2.I.69, 24.VIII.358, 2.XII. 362, 16.VIII.555, 25.V.1719, 15.IV.1878, 10.VII.1894); Erzincan has suffered similarly twenty-six times (in 1011, 1045, 1161, 1165, 1166, 1168, 1236, 1251, 1254, 1268, 1281, 1287, 1289, 1290, 1308, 1374, 1422, 1433, 1458, 1482, 1543, 1578, 1584, 1605, 1784, 1939). The Turkish government is now attempting to rebuild the town outside the fault zone and in accordance with the principles of modern earthquake construction developed in Japan and the Dutch East Indies. Erzurum was either completely or in greater part destroyed at least thirteen times (in 1268, 1458, 1482, 1584, 1659, 1794, 1843, 1850, 1852, 1860, 1861, 1901, 1902). Kars (near the Russian frontier), Lenihakan, and Erivan (both in Russia near the Alagözdağ) have been demolished several times, as they lie in the neighborhood of the Aras graben.

In western Turkey, south of the Izmit-Erzurum line, several nearly east-west grabens are probably due to north-south stretch phenomena in the curved part between the Anatolian and Greek orogens. These will be discussed later in greater detail. Here it is sufficient to state that many earthquakes originating in these grabens (the Iznik, Bursa, Endremit, Bergama, Gediz, Izmit, and Kücük Menderes grabens, and the Büyük Menderes graben north of Rhodos) indicate that likewise in this western part of Turkey the upfolding phase of the Anatolian orogenic belt has not yet ended.[6]

The Anatolian orogenic belt is cut near the southern boundary by the north-south Dead Sea graben zone, which continues through Syria, into Turkey

[6] Heavy earthquakes at Dikili in the Bergama graben on 22.IX.1939; destruction of Bergama in 1296; of Izmir in 178 B.C., and A.D. 1056, 1668, etc.

east of Antakya (Antioche), and east of the Gavurdağ as far as Maraş. This zone is also still active, as is proved by destructive earthquakes in Antakya in 713, 1091, and 1156 and near Maras in 1114.

Unquestionably, the Izmit-Erzurum line is a fault zone of regional importance. It also forms, roughly, the boundary between two different areas, the so-called Pontic orogen to the north and the the central Anatolian plateau (= Anatolides) south of the fault. The different stratigraphic development of the two units can be seen in Table III.

THE CENTRAL ANATOLIAN PLATEAU (ANATOLIDES)

The central Anatolian plateau is a very rigid zone, mainly composed of Paleozoic sediments, schists, and igneous rocks and covered by not very strongly folded Mesozoic sediments. These Anatolides of P. Arni were consolidated during the Hercynian folding, but they subsided again during the Mesozoic. Although the subsidence began in the Triassic, large parts of these high Hercynian mountains were not covered by the sea until late Mesozoic time. The Alpine uplift started first in these central parts. Owing to these two factors, the thinly developed Mesozoic cover was more eroded on this central Anatolian plateau than in the Toros Mountains to the south, and it disappeared completely farther west. During the Tertiary upfolding of the Anatolian orogenic belt the crestal part of this big dome broke and subsided with respect to the flanks. This became the site of a large inland sea during the Oligocene and Miocene, as is evidenced by lagoonal deposits and lignites.

THE PONTIDES

The northern flank of the Anatolian orogenic belt, i.e., the belt north of the plateau, is formed by the Pontides, which are composed of two geanticlinal strips separated by a trough. The southern flank (south of the plateau) is formed by the Torides or Toros orogen. On both sides of the Anatolian orogenic belt we find a foredeep—the Mesopotamian (previously discussed) on the south and the Pontic foredeep, containing the oil fields of the Kuban depression, on the north. The stratigraphic development of the Pontides is treated in Tables III and V.

THE TOROS OROGEN

Between the Toros orogen and the central Anatolian plateau lies a zone with several large volcanoes, some of which were still active in historic times. This volcanic inner zone can be compared with the volcanic inner zone of the Malayan orogen.

MAIN DIRECTIONS OF THRUSTING

The Toros orogen is a very complicated group of mountains with much variation in their stratigraphic and structural features. Thrusting both toward the north and toward the south can be observed. Both were probably caused by movements of the Arabian shield toward the north. If these movements led to thrusting at a high level in the earth's crust, we should find displacements in the younger sediments directed toward the north; but if the compression was more effective at a greater depth, this underthrusting would turn the younger sediments toward the south. Which manifestation took place was determined by the structure of the Arabian shield, the varying resistance in front of the shield, and other factors. The generalization often put forward, particularly by German authors, that the Pontides are folded only toward the north and the

Torides only toward the south is definitely incorrect.

DIFFERENCES IN TIME OF INTRUSION

Also differences in time of serpentine intrusion in the Torides can be observed both in east-west and north-south directions (see section on "Divisions and Their Classification, Toros Orogen"). In the Bitlis-Hakkari region (eastern Turkey) the main intrusion took place at the end of the Turonian and the beginning of the Senonian, whereas immediately south in the Siirt-Beytuşşebap zone no intrusion at all took place in this period. These large-scale serpentine intrusions, which are so characteristic of the Toros orogen, represent the parent-rock for the rich chromium deposits of Turkey.

FOLDING PERIODS

In the central Toros geanticline the first large-scale block movements took place at the end of the Turonian, when most of this area came above sea-level and delivered sediments for the Senonian and Lower Eocene formations of the southern areas, where subsidence continued until the end of the Lower Eocene. Both areas were strongly uplifted in the Middle Eocene, followed by a further subsidence of the southern areas during the Neogene. Most of the Torides, however, remained above sea-level after the Middle Eocene. It was only in the troughs between the central and southern Toros geanticlines that an extremely thick Neogene section could be deposited (see Table II, "Seyhan Toros"); this is one of the thickest known in the world and is not due to duplication of sections by faulting.

The Torides present many other interesting features, but a fuller discussion cannot be given in this article.

POSSIBLE CONTINUATION OF THE ANATOLIAN OROGENIC BELT IN SOUTHEASTERN EUROPE

CONTINUATION OF ANATOLIAN TECTONIC UNITS IN THE BALKANS

Kober, Kossmat, Renz, Philippson, Von Seidlitz, and others have tried to give a geotectonic analysis of the Balkans. However, they have had to face the fact that the geology of the western Balkans is still very poorly known. The author also knows the Balkans, except for a tourist trip, only from studying published articles, thus making any comparison with Turkey largely conjectural. But there are four structural elements or features in Greece which make a comparison between Turkey and Greece still possible. They are as follows:

1. The continuation of the Izmit-Erzurum line
2. The lithology of the Cyclades
3. The situation of the active Santorini volcano and the extinct volcano of Milos
4. The presence of the so-called "eastern Hellenic" and "Vardar" zones of Von Seidlitz in eastern Greece

The westerly continuation of the Izmit-Erzurum fault line has already been considered. As far as the Chalkidikean Peninsula in Greece, there seems to be little doubt about the situation of the fault zone, but whether it enters the Struma Valley or not is an open question.

In eastern Greece and Yugoslavia there are two zones described by Von Seidlitz—the eastern Hellenic zone and the Vardar zone. The eastern Hellenic zone (in the eastern Peloponnesus) is, according to the literature, composed of Paleozoic rocks covered by Triassic and Jurassic beds (with serpentines, gabbro, and diabasic intrusions), which are unconformably overlain by Upper Creta-

ceous strata. The Vardar zone, which lies east of the eastern Hellenic zone, is apparently composed of Paleozoic graywackes, marbles, etc., covered by Triassic and Jurassic beds of abyssic facies with many large basic intrusions (serpentine, etc.), unconformably covered by Upper Cretaceous flysch. P. de Wÿckerslooth[7] has indicated on a map the different chromium deposits of Greece and Yugoslavia, all of which lie in this Vardar zone, with large serpentine intrusions. There seems to be very little doubt that this Vardar zone can be compared with our central Toros geanticline, which has similar large-scale serpentine intrusions, rich chromium deposits, and a similar lithologic development of the Mesozoic.

The Cyclades are composed of metamorphic rocks like those of western Turkey. If we consider the Cyclades as the continuation of the Anatolides and try to connect the Torides via Crete, south of the Cyclades, with eastern Greece, we get a rather simple regional fold bending around a vertical axis.

In the core below the anticlinal crest lie the Cyclades. Between the Cyclades and the continuation of the Torides we find the volcanoes of Milos and Thera (Santorini). This arrangement can be compared with our southern volcanic zone in Turkey between the Torides and the central Anatolian plateau.

The previous discussions make it most plausible that the western continuation of the Izmit-Erzurum line must be east of the Vardar zone (being the continuation of the Toros geanticline) and, according to the serpentine and chromium-ore distribution in Yugoslavia, it should be east of a line which begins near the point of confluence of the Drina and Save, and extends east of the Kopaonik Mountain (west of the Morova Valley), east of Pristina and Skolpje, east of the Vardar Valley, and through the Gulf of Kassandra (western Chalkidike).

East of this Vardar zone in Bulgaria begins the so-called "Rhodope Massif," mainly composed of marbles, schists, and other Paleozoic rocks, similar to western and northwestern Turkey. All these facts make it reasonable to continue our Izmit-Erzurum line through the Rhodope massif, along the Struma and Morava valleys. By adopting this line we may recognize and compare the following zones in the Balkans:

1. The Dalmatian cost and western Peloponnesus, which should represent the southern Toros geanticline.
2. Central Yugoslavia and eastern Greece, representing the central Toros geanticline.
3. The volcanoes of Milos and Thera, representing the southern volcanic zone.
4. The Cyclades and the areas in Greece and Yugoslavia between the Vardar zone and the Morava-Struma Valley should represent the western thinning of the central Anatolian plateau, similar to the eastward thinning east of Sivas in Turkey.
5. The Struma-Morava depression, expressing the Izmit-Erzurum fault zone.
6. The Rhodope massif, comparable to the western crystalline part of the most southern zone of the southern Pontic geanticline, probably continuing with the Balkan Mountains of Bulgaria into the Transylvanian Alps of Rumania.

CONTINUATION OF BALKAN TECTONIC UNITS IN NORTHERN ITALY

The next question is whether we can accept Salomon-Calvi's assumption that

[7] "Die Chromerzprovinzen der Türkei und des Balkans und ihr Verhalten zur Grosstektonik dieser Laender," *Maden Tetkik ve Arama Enstitüsü Publication, Bull. No. 7* (1942).

the Izmit-Erzurum line is the same as his Tonal line of northern Italy. In order to study this possibility, we must summarize the main characteristics of this fault line in the southern Alps.

Proceeding southward from the areas near the Swiss-Italian frontier, which contain the "roots" of the Pennine nappes, we can distinguish the following zones:

1. *Tonal metamorphic zone.*—The Tonal zone comprises phyllites, garnet-staurolite schists, with many marble and amphibolite intercalations. In the western part of this unit large tonalite intrusions occur (Disgrazia massif, etc.).

2. *Insubrian fault zone of A. Spitz* (1919) *or Tonal line of Salomon-Calvi* (1890).—The fault zone runs from Giubiasco (south of Bellinzona, southeastern Switzerland) in an east-west direction north of the Val Tellina (northern Italy) and south of the Val Camonica. Farther east near Dinaro the fault is displaced toward the north by a cross-fault (the "Judicarian line"). The Insubrian fault zone continues farther eastward north of the granite massif of Brixen as the so-called "Puster line," which extends farther east into the "Drauzug" at least as far as the "Karawanken Mountains."

From Giubiasco the Insubrian fault continues westward into the "Cento Valli" and the "Zone du Canavèse," thence turning toward the southwest and following the east side of the "Sesia Lanzo zone" (root zone of the Pennine nappes). West of Lake Como (northern Italy) the fault zone is more commonly known as the "Jorio line of R. Staub"; east of Tirano it is known as the "Tonal line" of Salomon-Calvi.

The Insubrian zone is characterized by much mylonitization, which is greatest on the north side. Between individual mylonitic zones may be seen wedges, composed of Triassic, Permian, and Upper Carboniferous rocks, which originally covered the crystalline rocks farther south. Pre-Permian granite intrusions (the "Gneiss chiari") along the Insubrian fault zone indicate that this was a line of weakness even in Paleozoic time.

3. *Orobic metamorphic zone.*—In the east (probably Paleozoic) phyllites dominate, with few marble intercalations; in the west garnet-staurolite gneisses and garnet-mica schists occur, with few marble but many amphibolite intercalations. These amphibolite intercalations increase westward and become abundant in the "Ivrea zone."

4. *Orobic fault zone of C. Porro.*—In the central part of the Lombardy Alps, the Orobic fault zone runs from Bellano (at Lake Como), south of the Corno Stella Mountains, northeast of Mount Venerocolo into the Oglio Valley near Lava. This northeast part is often called the "Gallinera line" of R. Staub. The fault zone breaks through the Adamello group (tonalite massif) and continues probably south of the Dolomites as the "Val Sugana line." West of Lake Como the fault runs from Acquaseria, via Sasso Grande, Manno, and Cremenaga (in the Tresa Valley) to Bedero Val Travaglia (near Lake Maggiore), crossing the lake to Meina. The Orobic fault zone also contains pre-Permian granite intrusions (the "Gneiss chiari") and, in addition, young intrusions (near Baveno, Adamello group, etc.) which have accompanied the Tertiary deformation.

5. *Orobic and Lugano Alps.*—In the central parts the Orobic Alps are mainly composed of strongly imbricated Permian and Triassic sediments; to the east and west the younger Mesozoic sediments dominate. The Permian is mainly a volcanic effusive series. The Triassic is developed as thick limestone units with

marly intercalations in the Werfenian, Carnic, and Rhaetic. The Liassic is composed of black flaggy limestones, with red marbles in the upper part; the Dogger is composed of yellow and white limestones with radiolarite intercalations; the Malm consists of reddish (also greenish) limestones with red and greenish chert intercalations; the Lower Cretaceous is developed as a white limestone section, which changes into a flysch series in the Upper Cretaceous. In other words, there is a great similarity in the stratigraphic development of the western part of the central Toros geanticline and in the Orobic Alps (see Table II, Seydişehir and Seyhan Toros). The only fundamental difference is the lack of large-scale basic intrusions in the Orobic Alps at the end of the Jurassic and during the early Senonian. In this respect there is a greater resemblance to the eastern part of the southern Toros geanticline.

Also the Dalmatian coastal area and the western Peloponnesus were apparently developed in the Orobian facies without serpentine intrusions; eastward in the Vardar zone there were large serpentine intrusions.

Western Yugoslavia and western Greece are therefore probably the continuation both of the southern Toros geanticline and of the Orobic Alps. A fault zone similar to the "Orobic fault zone" occurs in eastern Turkey between the Siirt-Beytuşşebap zone (southern Toros geanticline) and the Bitlis-Hakkari Toros (central Toros geanticline). In addition, the northeast-southwest-directed blocks in the coastal plain southeast of Adana, near the Gulf of Iskenderun, indicate regional faulting between the central and the southern Toros geanticlines. In Greece and Yugoslavia the contact between the Vardar zone in western Greece seems to be an abnormal fault contact with large overthrusts similar to the Orobic faults.

If we consider this "Orobic fault zone" to be a regional one, our previous assumption about a continuation of the central Toros geanticline and Anatolides into the Vardar zone and western Rhodope massif seems to be correct. Both zones probably continue into the "Orobic metamphoric zone" (third tectonic unit) of the Lombardic Alps, which, however, is much thinner than its continuation in the Balkans. Also in the western part of the "Orobic metamorphic zone" (the so-called "Ivrea zone") many large amphibolitic intrusions penetrate the Paleozoic schists, comparable with the serpentine intrusions of the central Toros geanticline.

The Tonal metamorphic zone, separated from the "Orobic metamorphic zone" by the Insubrian fault zone, can be compared with the eastern Rhodope massif of Bulgaria. According to this interpretation, Salomon-Calvi's assumption seems to be correct. The Insubrian fault zone might continue, indeed, from the Karawanken in Austria toward the Morava-Struma Valley in the Balkans. The thinning of the different zones north and south of this fault zone toward eastern Turkey and the northwestern Balkans is in agreement with the phenomena to be expected in a regional fold curving around an area of alternating plastic and brittle layers (so-called "exfoliating leaf-shaped folding of Seidl," explained in the writer's work on the mechanism of the geological undulation phenomena, etc. [Leyden: Sythoff Pub. Co., 1937], p. 25).

In the Sesia-Lanzo zone (northwestern Italy) and in the areas north of the Tonal metamorphic zone are located the roots of the upper Pennine nappes and of the lowest eastern Alpine nappes. Between these regions are large Eocene tonalite

intrusions (e.g., The Bergell-Disgrazia massif). In the writer's publication[8] the retrofolding hypothesis of Staub has been shown to be incorrect. The upper eastern Alpine nappes are without roots; they have only a *pays d'origine* near the huge tonalite intrusions. The rising basement blocks, which developed the Pennine nappes, derooted the eastern Alpine crystalline blocks and transported them northward. Large intrusions took place along the scar, which closed later because of the further narrowing of the geosyncline (for further details see pp. 175–77 of ftns. 7 and 8).

It is also northeast of the Rhodope crystalline zone in northeastern Bulgaria that large overthrusts were reported by E. Boncev.[9] They probably continue into the Rumanian Transylvanian Alps with their large overthrusts. If we continue the Transylvanian Alps into the Carpathian Mountains with its nappe structures and finally into the Swiss Alps, we must assume that the rigid zone between the Pontides and the Torides widens again in northern Yugoslavia and forms the Hungarian plain west and east of the nappe zone.

We assumed that the Karawanken fault zone continues into the Morava-Struma zone. This transition (if our assumption is correct) is probably not a simple, regular one. Instead, it is probably cut by large north-south cross-faults.

Our present analysis cannot be complete for lack of sufficient trustworthy geological data on the Balkans. But with the aid of the well-known areas in the southern Alps (systematically studies by Leyden University parties, under the leadership of Dr. B. G. Escher) and in western Turkey (studied by the M.T.A. Enstitüsü), we have been able to correlate the zones of these regions in accordance with the facts available. We realize, however, that this synthesis is still very hypothetical and therefore have not attached a geotectonic map to this article.

[8] *La Géologie du Valle del Bitto et la tectonique des Alpes lombardes*, pp. 255–58.

[9] "Aus der alpidischen Tektonik Bulgariens," *Zeitschr. f. bulg. geol. Gesellsch.*, Vol. XIII (1940).

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REVIEWS

Geomorphology: An Introduction to the Study of Landforms. By C. A. COTTON. 4th ed. New York: John Wiley & Sons, Inc., 1947; printed by Whitcombe & Tombs, Ltd., New Zealand. Pp. xii+505; figs. 473. $6.00.

The new fourth edition of this well-known text appears for the first time in an American edition. Revisions from the third edition, which appeared in 1942, are minor: About twenty new and improved photographs have been substituted for older illustrations; slight changes have been made in the text; and the reading references and Index have been brought up to date and enlarged. It remains one of the most lucid discussions of geomorphology available for an introductory course.

LELAND HORBERG

Geology and Ground-Water Resources of the Island of Hawaii. By H. T. STEARNS and G. A. MACDONALD. (Territory of Hawaii, Division of Hydrography, Bull. 9.) Prepared in cooperation with U.S. Geological Survey. Honolulu, 1946. Pp. xiii+363; pls. 54; figs. 60

Bulletin 7 of this series, *Geology and Ground-Water Resources of the Island of Maui, Hawaii,* was reviewed in the *Journal,* Volume LIII (1945), page 144. Bulletin 8, *Geology of the Hawaiian Islands,* by H. T. Stearns, appeared early in 1946. This is a shorter synopsis of the geology of the whole group of Hawaiian Islands and may be recommended for its broader scope. It is effectively and beautifully illustrated, some of the plates being in color, while the numerous geologic maps are introduced as text figures where they are most serviceable.

Hawaii, largest of the Hawaiian Islands, has been built up by lavas poured from five volcanoes, each with independent rift zones and individual geologic history. Consequently, the first half of Bulletin 9 is devoted to the geomorphology, geologic features, and history of these volcanoes—Mauna Loa, Kilauea, Hualalai, Mauna Kea, and Kohala—intensive studies of which have already contributed greatly to the science of volcanology. The concise but comprehensive summary treatment of each of these brings together in readily available form much valuable information. In fact, this is the only recent publication which presents a compilation of all that is known of the geology of the island of Hawaii. Incidentally, the authors bring out the fact that volcanism in Hawaii "is waning in common with volcanism elsewhere on earth."

An excellent chapter on the "Petrography of Hawaii," by Macdonald, and one hundred and fifty pages on the ground-water resources complete the volume.

R. T. C.

The Coastline of England and Wales. By J. A. STEERS. Cambridge: University Press, 1946. (Macmillan Co. agents for United States and Canada.) Pp. xvii+644; figs. 114; pls. 115+2 pls. in natural color. $9.50.

Although the material for this study was prepared for the British Ministry of Town and Country Planning, the book is strictly physiographical and in no sense a treatise on land use. The approach is primarily descriptive, i.e., physiographic rather than geomorphic.

An introductory section is followed by a chapter on the structure and physiography of England and Wales, which includes a comprehensive appendix on stratigraphical geology. This is followed by a general summary of shoreline processes. Beach drifting by wave action is considered much more important than currents in transporting materials. The next eight chapters form the main part of the book and contain detailed descriptions of the entire coastline. In the final three chapters the problems of Pleistocene and Recent vertical movements, coastal dunes, and salt marshes are discussed. Geologists will find the opening and concluding chapters particulary useful. The descriptive chapters will be of interest largely to travelers and those concerned with special areas.

LELAND HORBERG

Geological Investigation of the Alluvial Valley of the Lower Mississippi River. By HAROLD N. FISK. Conducted for the Mississippi River Commission, Vicksburg, Miss., December 1, 1944; War Department, Corps of Engineers,

U.S. Army. 4to; pp. 78; atlas, 33 pls., 11 tables, 80 figs.

This superb work of which, it is understood, only one hundred copies were printed is monumental in its importance in the field of geology. Many of the plates are colored maps in great detail, a square yard or more in area, and the data recorded on them not alone are from surface observations but include those from many thousands of deep-well borings. The study represents the labors of a large staff, both engineers and geologists, carried out mainly during four recent years. The practical aim has been the greater security of the inhabitants of the flood-plain-delta region; and, as a result of new revetments on levees, some four million additional farmers have settled on the fertile plains.

The alluvial valley of the Mississippi, that below Cape Girardeau, Missouri, is a "valley within-valley," alluviated during the Pleistocene. During the earliest of the Pleistocene glaciations sea-level was several hundred feet lower than now, and the river channels had become deeply incised within the Mississippi embayment of the Coastal Plain, which extended as far north as Cape Girardeau. The resulting topography, now buried under the alluvium, is therefore rugged, and the bottoms of the trenches are in many places farther below the Tertiary plain than the exposed valley walls rise above it.

The valley outline is that of an elongated cornucopia, narrower from the Gulf to Natchez in Mississippi but widening to an open end at the north, with an almost straight northwest marginal contact with the Ouachita and Ozark highlands in Arkansas and Missouri.

The northern part of the valley has an average width of 70 miles and an elevation at the northern end of 350 feet and of 200 feet at the southern. All major tributaries except the Ohio enter from the west. The embayment is directly over an axis of downwarping throughout the northern part. This can be traced as a sinuous line southward as far as New Orleans, where it intersects the axis of an east-west trending downwarp—the Gulf Coast geosyncline.

A very remarkable structural fracture pattern is clearly revealed by the orientation of streams, both tributary and distributary. To quote Fisk:

> Those of the western tributaries, such as the White, Arkansas, and Red rivers, trend northwest-southeast; those of the eastern tributaries, such as Homochitto, Big Black and Yaluhusha rivers, trend northeast-southwest. These directions parallel the trends of the major fault zones of the region and show that the location of the tributary streams has been controlled by the regional structures.

The number of the faults of this rectangular region pattern is legion, and they are proved individually by the offsets revealed by many thousands of closely spaced deep-well borings which intersect the contact with the indurated rocks at the bottom of the embayment. There is probably no area anywhere in the world where deep drillings have been so closely spaced. Movements on these faults have occurred concurrently with the crustal subsidence under the weight of the accumulating delta sediments. That such movements are still continuing seems to be shown by the fault which appeared at the surface at Vacherio, Louisiana, in April, 1943, as well as by the fault escarpments which still remain from the great earthquake of New Madrid in 1811–12. A direct correlation is drawn also between the warping of the fracture system and the changes in trend of the successive deltaic increments.

The later history of the valley, that of alluviation since the entrenchment of the floor, is comprised in two major stages, the earlier recorded in the subsurface alluvium. The history of the later of these is marked by an alternation of rising and stationary sea-level with a lowering one. This history is recorded in a nest of four trenches, each bordered by a pair of terraces. These are found both in the Mississippi Valley and in those of its principal tributaries.

While the glacier of the first stage (the Nebraskan) was over a considerable part of North America, the sea-level was lowered and the Mississippi cut down its channel. During the deglaciation which followed, sea-level rose and alluviation produced the highest pair of terraces, the Williana terrace. The three succeeding glaciations (the Kansan, Illinoian, and Wisconsin) each brought about a lowering of sea-level, followed by a rise and heavy alluviation during deglaciation. These three lower terraces Fisk has designated the Bentley, Montgomery, and Prairie terraces, each with gravels at the base and grading upward through sands into clays at the surface.

Throughout this history the land to the northward had been upwarped. The maximum altitude of the Williana terrace is 350 feet, that of the Bentley 200 feet, of the Montgomery 100 feet, and of the Prairie 40 feet above the present floodplain of the river. All four terrace pairs slope gulfward; the uppermost, the Williana,

with the steepest gradient, the lowest, the Prairie, with the flattest. This is the convincing proof of their succession in age from top to bottom. This reviewer has had the prized opportunity of visiting the terrace series under the guidance of the author of the monograph.

WILLIAM HERBERT HOBBS

The Voyage of Captain Bellingshausen in the Antarctic Seas 1819–1821. Edited by FRANK DEBENHAM. Translated from the Russian. Hakluyt Society, 2d ser., Vol. XCI, pp. xxx +213. London: printed in Great Britain by W. Lewis; Cambridge: University Press, 1945. $15.

The appearance of this translation a full century and a quarter after the completion of the voyage is an event of importance in the history of polar exploration. The Russian original was not published until a decade after the return of the expedition. No copy of the original report had been discovered in this country and only two in Great Britain. A copy of the atlas is, however, in the Library of Congress. Two partial translations had appeared in Germany.

Bellingshausen's expedition was one of the greatest, as well as one of the earliest, to penetrate into Antarctic seas, and his discovery of Antarctic land in latitude 69° S. (Peter I Island) was antedated by but one Antarctic landfall, that by the American captain, Nathaniel B. Palmer, only thirty-three days earlier. The discovery of Palmer Land, now the Palmer Peninsula of the Antarctic Continent, was made on November 18, 1820, while Bellingshausen sighted Peter I Island on January 9, 1821. On February 6 Palmer's little sloop "Hero" had an accidental joining-up off Palmer Land with Bellingshausen's "Vostok" and "Mirnyi." The two captains had a meeting on the "Vostok," where they conversed through an interpreter. We now have the accounts of this conversation by both parties to it.[1] They are in perfect accord, though that of the Russian captain occupies less than a page of his narrative, while Palmer's story is much more complete. Palmer states that Bellingshausen appeared to be astounded when he displayed his chart and his log, and the Russian exclaimed: " 'What do I see and what do I hear from a boy in his teens—that he is commander of a tiny boat of the size of a launch of my frigate . . . . and sought the point I, in command of one of the best appointed fleets at the disposal of my august master, have for three long, weary, anxious years searched day and night for.' With his hand on my head he added, 'What shall I say to my master? What will he think of me? But be that as it may, my grief is your joy; wear your laurels with my sincere prayers for your welfare. I name the land you have discovered in honor of yourself, noble boy, Palmer's Land.' "

This part of the conversation is omitted in Bellingshausen's account, and Debenham, the editor of the translation, has devoted four of the twenty pages of his introduction to an acrimonious argument in an attempt to discredit Palmer as the discoverer of the Sixth Continent.

Priority in discovery is, of course, not to be settled by this report. Bellingshausen with little doubt regretted his impetuous speech, as he reflected that the right to name new lands belongs to the discoverer, and he knew he must win the approval of his august master.

That Palmer's Land was actually discovered on November 18, 1820, is proved by Palmer's log, now in the Library of Congress; also by the appearance of "Palmer's Land" on an American map of September 28, 1821 (see *Science*, Vol. ooo [June 23, 1939], pp. ooo–oo), and on official British Admiralty maps of November 1, 1822, and November 4, 1824. It was later that the Admiralty set up the claim that the same land had been discovered by Edward Bransfield, R.N., and named Trinity Land almost a year earlier than Palmer's discovery.

We learn from the translation that Bellingshausen's two sloops of war were well found and outfitted but that the captain was disappointed in failing to secure scientific personnel, especially naturalists. We have, however, the assurance of a distinguished ornithologist that Bellingshausen's own observations within that field were excellent and often brilliant.

An ardent admirer of Captain James Cook, Bellingshausen endeavored to make his own cruise supplementary. Where Cook had been forced by the pack ice to lay his course to the north, Bellingshausen made special efforts to maintain a more southerly cruise, in which he was generally successful. His meteorological observations were very valuable, and his determinations of position, checked by several observers, command the fullest confidence. The translation appears to have been well done, and the narrative is easy to follow.

WILLIAM HERBERT HOBBS

[1] For Palmer's account see "The Discovery of Antarctica within the American Sector as Revealed by Maps and Documents," *Trans. Amer. Phil. Soc.*, N.S., Vol. XXXI (January, 1939), pp. 18–23.

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THE SYSTEM $CaAl_2Si_2O_8$–$Ca_2Al_2SiO_7$–$NaAlSiO_4$

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ABSTRACT

The thermal-equilibrium relationships in the system $CaAl_2Si_2O_8$–$Ca_2Al_2SiO_7$–$NaAlSiO_4$ have been experimentally investigated by the quenching method. The data are tabulated, and the results are presented in equilibrium diagrams. Complex solid solutions in all the crystalline phases render the system nonternary throughout, and courses of crystallization must be viewed in terms of the quaternary system of the four component oxides. The composition of gehlenite mix-crystals obtained in this investigation is best expressed by assuming solid solution with a hitherto unknown "molecule," of the composition $CaO \cdot 2Al_2O_3$.

INTRODUCTION

This investigation is concerned with the thermal-equilibrium relationships, at atmospheric pressure, of the three silicates, $CaAl_2Si_2O_8$, $Ca_2Al_2SiO_7$, and $NaAlSiO_4$. These compounds are the principal "molecules" of the natural minerals, anorthite, gehlenite, and nepheline, respectively. Anorthite, as an end-member of the plagioclase feldspars, is of prime importance in igneous rocks. Nepheline, a feldspathoid, is the most characteristic mineral of the alkaline igneous rocks, while melilites, of which gehlenite is an end-member, are also feldspathoids that are present in alkaline rocks which are rich in lime. Nepheline and melilite are found associated in some igneous rocks, and gehlenite also occurs as a metamorphic mineral.

The only compound involved in this study that exhibits polymorphism is $NaAlSiO_4$; but carnegieite, the high-temperature modification, has never been found in nature. In addition to the compounds mentioned, the substance β- alumina was encountered and is of petrologic interest, inasmuch as corundum is a typical accessory mineral of the alkaline syenites.

The system constitutes a portion of a plane within the quaternary system soda-lime-alumina-silica. This quaternary system is of major importance in the physicochemical aspects of igneous petrology, as the constituent oxides form more than 80 per cent of the average oxide composition of igneous rocks. The relation of the system studied to the general quaternary system is shown in Figure 1. The system anorthite-gehlenite-nepheline forms one face of a subsidiary tetrahedron anorthite-gehlenite-wollastonite-nepheline within the quaternary system; and the plane anorthite-nepheline-wollastonite is common to a second subsidiary tetrahedron, anorthite-nepheline-wollastonite-albite.

Forty-six mixtures were prepared and subjected to thermal study. Three of

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these compositions were in the plane of anorthite-gehlenite-nepheline + 10 per cent wollastonite, being used to locate the position of the boundary between the fields of anorthite, melilite, and nepheline as it passes through this plane of 10 per cent $CaSiO_3$ in the subsidiary tetrahedron. In addition, five mixtures prepared and examined by H. S. Yoder[1] that fall in the plane AN-GEH-NE are incorporated with the results here presented.

The mixtures were studied as to thermal behavior by means of the quenching method developed in the Geophysical Laboratory of the Carnegie Institution of Washington.[2] Temperatures were measured with a platinum-platinum 90 rhodium 10 thermoelement, calibrated at the melting-points of lithium metasilicate (1,201° C.) and diopside (1,391.5° C.). The results of quenching runs were examined microscopically to determine the nature of the phases present at equilibrium. The significant data are presented in Tables 1 and 2, and from these data the equilibrium diagram has been prepared.

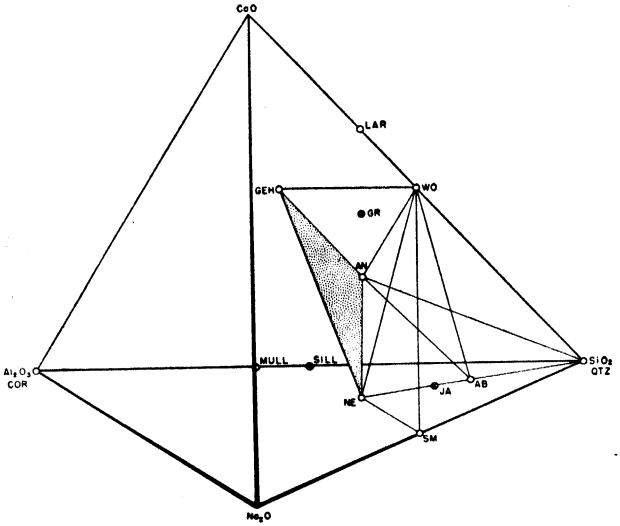

FIG. 1.—The soda-lime-alumina-silica tetrahedron. The shaded area indicates the position of the system anorthite-gehlenite-nepheline within the tetrahedron. Other internal systems that have been investigated are shown by full lines joining the composition points. The plane nepheline-gehlenite-wollastonite plus 10 per cent anorthite has also been investigated (unpublished data of H. S. Yoder). The compounds that have been prepared synthetically are shown as ○; those that occur in nature but that have not been synthesized are shown thus: ⊗.

The abbreviations signify: *AN*, anorthite; *GEH*, gehlenite; *NE*, nepheline; *WO*, wollastonite; *AB*, albite; *JA*, jadeite; *SM*, sodium metasilicate; *SILL*, sillimanite; *MULL*, mullite; *LAR*, larnite; *GR*, grossularite; *QTZ*, quartz; *COR*, corundum.

PROPERTIES OF THE CRYSTALLINE PHASES

$CaAl_2Si_2O_8$

Pure $CaAl_2Si_2O_8$ has but one modification—anorthite—which melts at

[1] Unpublished data on the system nepheline-gehlenite-wollastonite plus 10 per cent anorthite.

[2] E. S. Shepherd, G. A. Rankin, and F. E. Wright, "The Binary Systems of Alumina with Silica, Lime, and Magnesia," *Amer. Jour. Sci.*, Vol. XXVIII (4th ser., 1909), p. 308.

1,552° C. It is triclinic, with $\alpha = 1.5755$, $\beta = 1.5832$, and $\gamma = 1.5885$. Crystals obtained in quenching runs are commonly platy, with 010 showing the strongest development. Albite twinning is common. The crystals encountered in this investigation are plagioclase, of a composition not far from pure anorthite, as shown by maximum index values approaching 1.588.

$Ca_2Al_2SiO_7$

Artificial gehlenite also melts congruently, at a temperature of 1,590° C.[3] It is tetragonal, as tabular or short prismatic crystals. Its indices of refraction are $\omega = 1.669$ and $\epsilon = 1.658$. The melilites encountered in this study were generally tabular, thus showing positive elongation on a negative crystal. Globular to shapeless forms were also encountered, particularly in compositions close to a boundary curve. The indices of melilites forming in this system were significantly lower than those of pure gehlenite. Although the indices were accurately determined in but several runs, values as low as $\omega = 1.651$ and $\epsilon' = 1.641$ were encountered. This will be discussed in more detail in a later section concerned with substances in solid solution.

$NaAlSiO_4$

Carnegieite, the high-temperature modification of $NaAlSiO_4$, exists in three forms, all of which are known only from synthetic melts. The highest-temperature form, which is isometric, melts at 1,526° C.[4] and inverts to a triclinic form at 690° C.[5] A second inversion, accompanied by a marked change in birefringence, occurs at 226.5° C. The indices of refraction of the low-temperature form are given by Bowen[6] as $\alpha = 1.514$, $\beta = 1.514$, $\gamma = 1.509$. Carnegieite usually occurs in globular masses, showing no external crystal form and, typically, has intricate polysynthetic twinning that forms as a result of inversion.

Pure artificial nepheline inverts to carnegieite at a temperature of 1,254 ± 5° C.[7] Bowen[8] found that anorthite, which may go into solid solution with nephelite to the extent of 35 per cent, raises the inversion temperature 100° C., while albite in solid solution brings about a rise of 30° at its maximum concentration.[9] Smalley, in investigating the system $NaAlSiO_4$–$Ca_2Al_2SiO_7$, found an inversion range extending from about 50° above, to 100° below, the normal inversion point.[10] However, the nonbinary nature of the system precluded making any assured statement as to the exact nature of the solid solution.

Nepheline is hexagonal and crystallizes from melts as prismatic crystals with basal pinacoid. The indices as determined by Bowen on the pure synthetic material are $\epsilon = 1.533$ and $\omega = 1.537$.[11] He found that the indices of the limiting anorthite-nepheline solid solution are

[3] E. F. Osborn and J. F. Schairer, "The Ternary System, Pseudo-wollastonite-Akermanite-Gehlenite," *Amer. Jour. Sci.*, Vol. CCXXXIX (1941), p. 719.

[4] N. L. Bowen, "The Binary System: $Na_2Al_2Si_2O_8$ (Nephelite, Carnegieite)–$CaAl_2Si_2O_8$ (Anorthite)," *Amer. Jour. Sci.*, Vol. XXXIII (4th ser., 1912), pp. 554 ff.

[5] N. L. Bowen and J. W. Greig, "The Crystalline Modifications of $NaAlSiO_4$," *Amer. Jour. Sci.*, Vol. X (5th ser., 1925), pp. 208–10.

[6] P. 564 of ftn. 4.

[7] J. W. Greig and Tom. F. W. Barth, "The System $Na_2O \cdot Al_2O_3 \cdot 2SiO_2$ (Nephelite, Carnegieite)–$Na_2O \cdot Al_2O_3 \cdot 6SiO_2$ (Albite)," *Amer. Jour. Sci.*, Vol. XXXV+A (5th ser., 1938) p. 108.

[8] P. 571 of ftn. 4.

[9] P. 109 of ftn. 7.

[10] Robert G. Smalley, "The System $NaAlSiO_4$–$Ca_2Al_2SiO_7$," *Jour. Geol.*, Vol. LV, No. 1 (January, 1947), p. 28.

[11] P. 566 of ftn. 4.

$\epsilon = 1.539$, $\omega = 1.537$, while solid solution of albite in nepheline has been found to lower the indices of the latter somewhat.[12] Unusually high indices were noted by Smalley in nephelines crystallizing from the system $NaAlSiO_4$–$Ca_2Al_2SiO_7$, values as high as 1.552 being reached in crystals from melts having 35 per cent gehlenite.[13] The present study indicates that the solid solution responsible for this rise remains effective upon addition of $CaAl_2Si_2O_8$, as values of $\omega = 1.548$ and $\epsilon > 1.548$ were obtained from runs at a composition of approximately 35 per cent nepheline, 20 per cent gehlenite, and 45 per cent anorthite.

β–Al_2O_3

The β–Al_2O_3 here encountered occurs in thin hexagonal plates, exhibiting high relief and high birefringence when viewed on edge. Parallel extinction and positive elongation are shown, the sign of elongation coupled with the platy habit indicating that the crystals are optically negative. No indices were determined, but G. A. Rankin and H. E. Merwin gave the values $\omega = 1.677 \pm .003$ and $\epsilon = 1.635$–1.650 for the β–Al_2O_3 found in preparations thought to be pure alumina.[14] Rankin and Merwin determined the melting-point as 2,050° ± 20° C.

The exact relations of β–Al_2O_3 to the other known forms of Al_2O_3 are not completely understood. The indices are considerably lower than those of corundum, and chemical and X-ray analyses indicate that soda is present in the structure. On the basis of X-ray work the formula $Na_2O \cdot 11Al_2O_3$ was proposed by C. A. Beevers and M. A. S. Ross,[15] whereas W. L. Bragg, C. Gottfried, and J. West had somewhat earlier determined it as $\frac{1}{2}\,Na_2O \cdot 11\frac{1}{2}Al_2O_3$.[16]

Recent work at the Geophysical Laboratory of the Carnegie Institution of Washington has shown that β–Al_2O_3 is converted to corundum when held at elevated temperatures for long periods,[17] indicating that β–Al_2O_3 is metastable. The melting-point as determined by Rankin and Merwin is almost exactly that of pure α–Al_2O_3, so it is likely that their crystals had undergone the conversion to corundum before melting. It is possible that β–Al_2O_3 is an example of a "stuffed lattice," in which Na atoms assume random interstitial positions in the network. Certain soda-bearing tridymites have been interpreted in this way by M. J. Buerger.[18] This explanation would account for the disagreements of different investigators, as it does not require any fixed ratio of soda to alumina. It is also consistent with the findings of the Geophysical Laboratory, as the Na atoms, given sufficient time, might tend to be "squeezed out" of the structure.

PREVIOUS INVESTIGATIONS

THE SYSTEM $CaAl_2Si_2O_8$–$NaAlSiO_4$

The system $CaAl_2Si_2O_8$–$NaAlSiO_4$ was investigated by Bowen,[19] and his equilibrium diagram is shown in Figure 2, with modifications based on work of

[12] W. R. Foster, "The System $NaAlSi_3O_8$–$CaSiO_3$–$NaAlSiO_4$," *Jour. Geol.*, Vol. L (1942), p. 157.

[13] P. 28 of ftn. 10.

[14] "The Ternary System CaO–Al_2O_3–MgO," *Jour. Amer. Chem. Soc.*, Vol. XXXVIII (1916), pp. 566–88.

[15] "The Crystal Structure of β–Al_2O_3," *Zeitschr. f. Kryst. u. Min.*, Vol. XCVII (1937), pp. 59–66.

[16] "The Structure of β–Al_2O_3," *Zeitschr. f. Kryst. u. Min.*, Vol. LXXVII (1931), pp. 255–74.

[17] J. F. Schairer and N. L. Bowen, "Melting Relations in the Systems Na_2O–Al_2O_3–SiO_2 and K_2O–Al_2O_3–SiO_2," *Amer. Jour. Sci.*, Vol. CCXLV (April, 1947) pp. 193–204.

[18] *Science*, Vol. XCV (1942), pp. 20–21.

[19] Pp. 551–73 of ftn. 4.

J. F. Schairer[20] and of W. K. Gummer.[21] This system exhibits partial solid solution between the two end-members, complicated by the carnegieite-nepheline inversion. It was, in fact, the first-found example in silicates of the combined phenomena of solid solution and enantiotropism theoretically proposed by Roozeboom. The eutectic was placed at 46 per cent anorthite by Bowen and at a temperature of 1,304° C. The inversion takes place at 1,352° C., a rise of 104° above the inversion in pure $NaAlSiO_4$, as a result of the solid solution of anorthite in nepheline to the extent of 35 weight per cent. Carnegieite takes but approximately 5 per cent anorthite in solid solution.

Bowen encountered "alumina" in some of his mixtures and accounted for it by soda volatilization at the high temperatures required for fusion of the glasses. This actually occurs but is only part of the story. The β-form was at that time unknown, but, in investigating the system $NaAlSiO_4$–$CaAl_2Si_2O_8$–SiO_2, Schairer[22] found β–Al_2O_3 in mixtures adjacent to the nepheline-anorthite join; and, on preparing mixtures along that join, he also found a considerable range in which β–Al_2O_3 is the primary phase. This work was somewhat amplified by Gummer,[23] who determined the extent of the field of β–Al_2O_3 in the system $CaSiO_3$–$CaAl_2Si_2O_8$–$NaAlSiO_4$. The liquidus for β–Al_2O_3 has not been definitely established as yet, but its outline, based on the evidence on hand, is shown by a dashed line on the diagram.

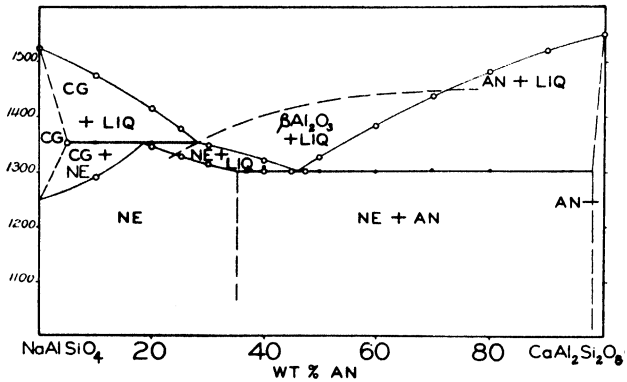

FIG. 2.—Equilibrium diagram for the system $CaAl_2Si_2O_8$–$NaAlSiO_4$ (after Bowen, with modifications after work by Schairer and Gummer).

This intrusion of a field of β–Al_2O_3 upsets the binary nature of the system. The field of nepheline plus liquid is actually a field of β–Al_2O_3, nepheline, and liquid; and much of the field of anorthite plus liquid, as determined by Bowen, is also a ternary field of β–Al_2O_3, anorthite, and liquid. The subliquidus extent of the field of β–Al_2O_3 has not been determined, but Gummer's work indicates that it is present in several mixtures at 1,200° C.

[20] Unpublished data.

[21] "The System $CaSiO_3$–$CaAl_2Si_2O_8$–$NaAlSiO_4$," *Jour. Geol.*, Vol. LI, No. 8 (November–December, 1943), pp. 508–9.

[22] Unpublished data.

[23] Pp. 508–9 of ftn. 21.

THE SYSTEM $NaAlSiO_4$–$Ca_2Al_2SiO_7$

The equilibrium diagram of this system is presented in Figure 3, as determined by Smalley.[24] The unusual nature of this system merits some discussion, which is here presented largely as taken from Smalley's work. In his words: "Even a casual inspection of the equilibrium diagram will suffice to show that this is not a true binary system."[25]

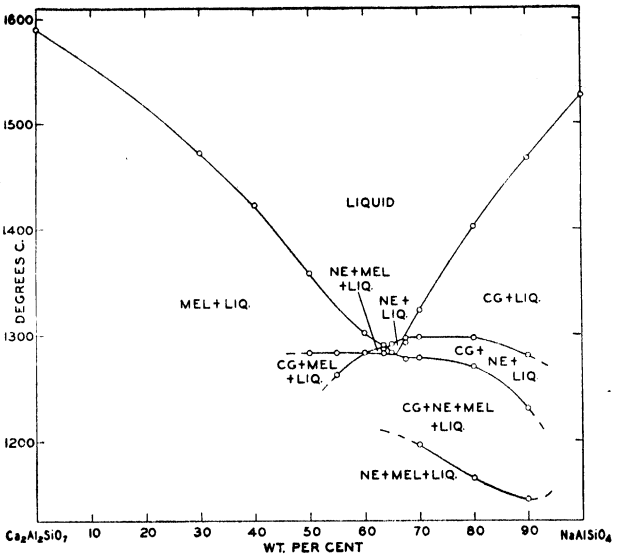

FIG. 3.—Equilibrium diagram for the system $NaAlSiO_4$–$Ca_2Al_2SiO_7$ (after Smalley)

Variations in optical properties of all three of the crystalline phases are indicative of solid solutions; that the solid solutions cannot be expressed in terms of the end-members is obvious from the diagram. The lack of a binary eutectic, of simple solidus curves and the persistence of the liquid phase down to the lowest temperatures of observation, as well as the previously mentioned large temperature range of inversion on both sides of the pseudoeutectic composition, all attest to the nonbinary nature of the system. The inversion relations are further complicated by the unique occurrence, near the pseudoeutectic, of nepheline in equilibrium with liquid at higher temperatures than is carnegieite.

In several fields of this system three phases occur in equilibrium, and in one field four phases are present. Consideration of the phase rule, $P + F = C + 1$, as applied to systems in which the vapor phase is negligible, indicates that in a binary system a maximum of three phases can coexist in equilibrium, and then only at a fixed temperature. Thus the existence of four phases within a temperature composition range requires four components—the four constituent oxides. The join $NaAlSiO_4$–$Ca_2Al_2SiO_7$ is therefore quaternary in nature, and the composition of the phases cannot be represented in the plane of the diagram but must be viewed in terms of the parent-tetrahedron.

The junction of the liquidus curves is

[24] Pp. 27–37 of ftn. 10.

[25] P. 31 of ftn. 10.

at 64 per cent $NaAlSiO_4$ at a temperature of 1,287° C. This pseudoeutectic is actually divariant, representing the trace of the boundary line between the fields of nepheline solid solution and melilite in the quaternary system. The junction of the carnegieite liquidus with the nepheline liquidus occurs at 67 per cent $NaAlSiO_4$ and at 1,297° C. Just as the pseudoeutectic is not invariant, so these liquidus curves are not univariant but trivariant, and the curves do not give the compositions of the liquids in equilibrium with the solid phases at the various temperatures.

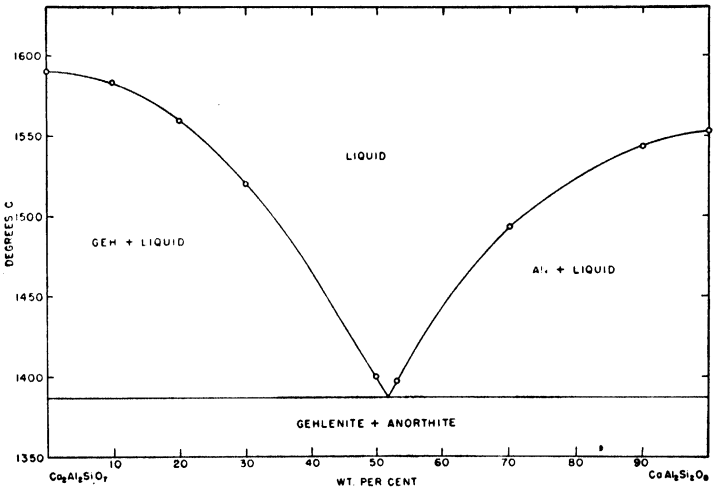

FIG. 4.—Equilibrium diagram for the binary system $CaAl_2Si_2O_8$–$Ca_2Al_2SiO_7$ (after Rankin and Wright, as revised by E. F. Osborn).

THE BINARY SYSTEM $CaAl_2Si_2O_8$–$Ca_2Al_2SiO_7$

The binary system $CaAl_2Si_2O_8$–$Ca_2Al_2SiO_7$ was first investigated by G. A. Rankin and F. E. Wright in their work on the system CaO–Al_2O_3–SiO_2.[26] The binary join has been revised by E. F. Osborn,[27] and his data are here used in constructing the equilibrium diagram of Figure 4. The system shows none of the complications of the preceding limiting systems. The eutectic is at 51.7 per cent anorthite and at a temperature of 1,387° ± 2° C.

THE THERMAL DATA

The thermal data necessary for establishing the equilibrium relationships in the system $CaAl_2Si_2O_8$–$Ca_2Al_2SiO_7$–$NaAlSiO_4$ are presented in Table 1, and the diagrams based on these results are shown in Figures 6 and 7. In Figure 6 are shown the isotherms at 50° intervals, while in Figure 7 the isotherms have been omitted, the stability fields have been labeled, and arrows have been inserted to indicate the directions of falling temperature. Figure 5 is a composition–refractive index[28] diagram of the homogeneous glasses prepared in this plane.

[26] "The Ternary System of CaO–Al_2O_3–SiO_2," *Amer. Jour. Sci.*, Vol. XXXIX (4th ser., 1915), pp. 1–79.

[27] Unpublished data.

[28] Sodium-vapor light was used for all refractive index measurements in this study.

NONTERNARY NATURE OF THE SYSTEM

The system is not a true ternary system; indeed, it is quaternary throughout, since a field of basic plagioclase is present rather than one of anorthite, a field of melilite instead of gehlenite, and carnegieite and nepheline also form solid solutions that cannot be expressed in terms of the three end-members. The field of β–Al_2O_3 encroaches upon the system as an extension of the quaternary field of alumina originating at the Al_2O_3 corner of the tetrahedron. The system is thus quite complex, the composition of any liquid upon crystallization does not remain in the plane defined by the three end-members; courses of crystallization can be discussed only in terms of the parent–soda-lime-alumina-silica system. A more complete discussion as to the nonternary behavior of mixtures in this plane will be presented in a following section on courses of crystallization.

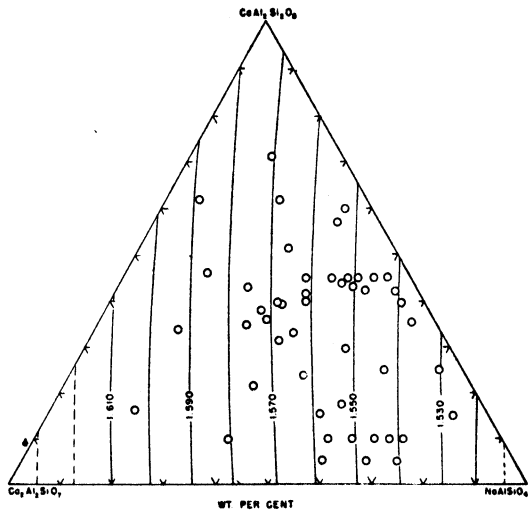

FIG. 5.—Composition–refractive-index diagram for the glass mixtures

THE LIQUIDUS RELATIONS

The five fields of the primary phases are, in order of decreasing size, melilite, plagioclase, nepheline solid solutions, carnegieite solid solutions, and β-alumina. The configuration of the fusion surfaces can be ascertained from the isotherms of Figure 6. The carnegieite surface dips rather sharply into the adjoining fusion surface of nepheline, which is relatively flat. The surfaces of melilite and plagioclase are convex and slope toward the center of the system. The fusion surface of β–Al_2O_3 is but tentatively indicated, owing to the somewhat uncertain data on the liquidus relations along the $NaAlSiO_4$–$CaAl_2Si_2O_8$ join.

A maximum is present on the boundary curve between the fields of melilite and nepheline solid solutions. This maximum, the position of which is indicated by the curvature of the 1,300° C. iso-

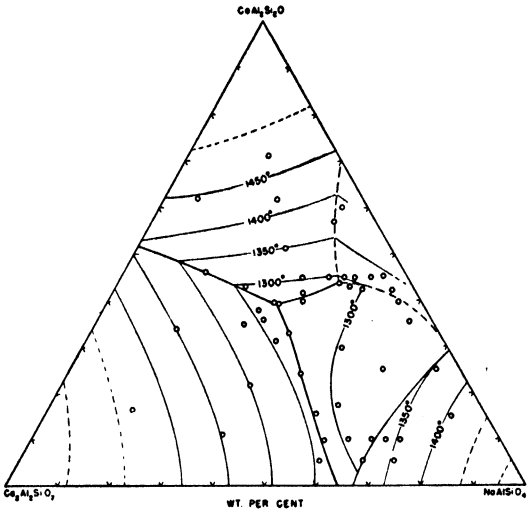

FIG. 6.—Equilibrium diagram for the system $CaAl_2Si_2O_8$–$Ca_2Al_2SiO_7$–$NaAlSiO_4$, with isotherms

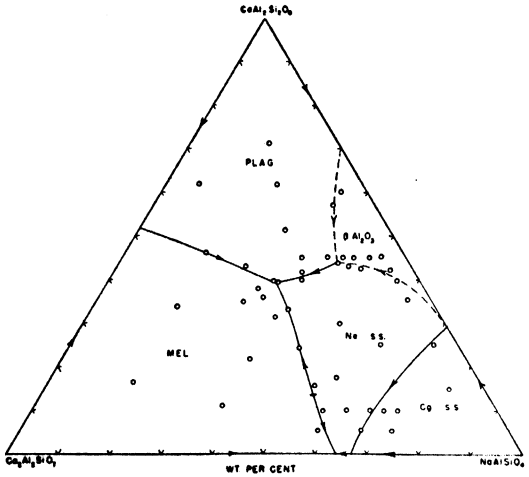

FIG. 7.—Equilibrium diagram for the system $CaAl_2Si_2O_8$–$Ca_2Al_2SiO_7$–$NaAlSiO_4$, with isotherms omitted

TABLE 1

THERMAL DATA FOR THE SYSTEM $CaAl_2Si_2O_8$–$Ca_2Al_2SiO_7$–$NaAlSiO_3$

Weight Per Cent			Time (Minutes)	Temperature (° C.)	Final Condition
$CaAl_2Si_2O_8$	$Ca_2Al_2SiO_7$	$NaAlSiO_4$			
Primary-Phase Melilite					
31.3	32	36.7	30	1299	All glass
			30	1295	Glass, rare MEL
5.2	36.8	58	30	1296	All glass
			30	1292	Glass, MEL
33.1	28.5	38.4	30	1278	All glass
			30	1274	Glass, NE, rare MEL
24	31	45	30	1288	All glass
			30	1284	Glass, NE, rare MEL
46	38.1	15.9	30	1328	All glass
			30	1324	Glass, MEL, rare AN
21.7	41.8	36.5	30	1352	All glass
			30	1348	Glass, MEL
11	52.6	36.4	30	1399	All glass
			30	1395	Glass, MEL
34	50	16	30	1407	All glass
			30	1403	Glass, rare MEL
36	32	32	30	1293	All glass
			30	1289	Glass, MEL
			30	1259	Glass, MEL, NE
			30	1242	Glass, MEL, NE, AN
38	32	30	30	1293	All glass
			30	1289	Glass, MEL
			30	1259	Glass, MEL, rare AN
			30	1255	Glass, MEL, AN, NE
35	36⅓	28⅔	30	1319	All glass
			30	1315	Glass, MEL
			30	1259	Glass, MEL
			30	1255	Glass, MEL, NE, AN
16.4	67.1	16.5	30	1491	All glass
			30	1486	Glass, MEL
Primary-Phase Anorthite					
71.3	13.4	15.3	30	1469	All glass
			30	1465	Glass and AN
39.7	28.1	32.2	30	1269	All glass
			30	1267	Glass, rare AN
			30	1264	Glass, AN, MEL
			30	1260	Glass, AN, MEL, NE

TABLE 1—*Continued*

Weight Per Cent			Time (Minutes)	Temperature (° C.)	Final Condition
$CaAl_2Si_2O_8$	$Ca_2Al_2SiO_7$	$NaAlSiO_4$			
Primary-Phase Anorthite—*Continued*					
45	20	35	30	1306	All glass
			30	1302	Glass, rare AN
			30	1259	Glass, AN, NE
			30	1251	Glass, AN, NE, MEL
45	15	40	30	1302	All glass
			30	1298	Glass, rare AN
			30	1288	Glass, AN
			30	1280	Glass, AN, NE
57	8	35	30	1376	All glass
			30	1372	Glass, rare β-Al_2O_3, AN
41.5	21.8	36.7	30	1283	All glass
			30	1279	Glass, rare AN
			30	1275	Glass, AN
			30	1271	Glass, AN, NE
			30	1250	Glass, AN, NE
			30	1242	Glass, AN, NE, MEL
46	38.1	15.9	30	1328	All glass
			30	1324	Glass, MEL, rare AN
43	32	25	30	1298	All glass
			30	1294	Glass, very rare AN
			30	1289	Glass, AN
			30	1285	Glass, AN, rare MEL
51.3	20.1	28.6	30	1352	All glass
			30	1348	Glass, AN
61.8	16.6	21.6	30	1422	All glass
			30	1416	Glass, AN
62	31.7	6.3	30	1449	All glass
			30	1445	Glass, AN
Primary-Phase Nepheline					
$15\frac{2}{3}$	32	$52\frac{2}{3}$	30	1295	All glass
			30	1291	Glass, NE
			30	1287	Glass, NE, MEL
33.1	28.5	38.4	30	1278	All glass
			30	1274	Glass, NE, rare MEL
24	31	45	30	1288	All glass
			30	1284	Glass, NE, rare MEL
42.4	9.9	47.7	30	1302	All glass
			30	1298	Glass, NE
43	12	45	30	1297	All glass
			30	1293	Glass, NE

TABLE 1—*Continued*

Weight Per Cent			Time (Minutes)	Temperature (° C.)	Final Condition
$CaAl_2Si_2O_8$	$Ca_2Al_2SiO_7$	$NaAlSiO_4$			
Primary-Phase Nepheline—*Continued*					
35.6	4.4	60	30	1329	All glass
			30	1325	Glass, NE
39.8	4.4	55.8	30	1319	All glass
			30	1315	Glass, NE
			30	1298	Glass, NE, rare β-Al_2O_3
5.2	28.4	66.4	30	1306	All glass
			30	1302	Glass, NE-CG
40	22.6	37.4	30	1273	All glass
			30	1269	Glass, NE
			30	1265	Glass, NE, AN
			30	1250	Glass, NE, AN
			30	1246	Some glass, NE, AN, MEL
17.4	27	55.6	30	1304	All glass
			30	1302	Glass, rare NE
29.7	20	50.3	30	1306	All glass
			30	1302	Glass, extremely rare NE
39⅓	27⅓	33⅓	30	1269	All glass
			30	1267	Glass, rare NE
			30	1263	Glass, NE, MEL
			30	1259	Glass, NE, MEL, rare AN
25.1	14.8	60.1	30	1327	All glass
			30	1323	Glass, NE
Primary-Phase Carnegieite					
5.1	22.6	72.3	30	1357	All glass
			30	1353	Glass, CG
25.1	4.4	70.5	30	1358	All glass
			30	1354	Glass, CG
5.2	28.4	66.4	30	1306	All glass
			30	1302	Glass, NE-CG
15.1	6.6	78.3	30	1406	All glass
			30	1402	Glass, abundant CG
Primary-Phase β-Al_2O_3					
45.2	4.4	50.4	30	1338	Glass, very rare β-Al_2O_3
			30	1342	All glass
45	7	48	30	1323	All glass
			30	1319	Glass, rare β-Al_2O_3
			30	1297	Glass, sparse β-Al_2O_3
			30	1280	Glass, NE, AN

TABLE 1—*Continued*

Weight Per Cent			Time (Minutes)	Temperature (° C.)	Final Condition
$CaAl_2Si_2O_8$	$Ca_2Al_2SiO_7$	$NaAlSiO_4$			
Primary-Phase β-Al_2O_3—*Continued*					
45	10	45	30 30	1302 1306	Glass, rare β-Al_2O_3 All glass
45	12	43	30 30	1306 1302	All glass Glass, rare β-Al_2O_3
60	5	35	30 30 30 30	1393 1389 1384 1376	All glass Glass, rare β-Al_2O_3 Glass, β-Al_2O_3 Glass, β-Al_2O_3, AN
43.9	13.6	42.5	30 30 30	1294 1292 1288	All glass Glass, very rare β-Al_2O_3 Glass, NE, AN, β-Al_2O_3
57	8	35	30 30	1376 1372	All glass Glass, rare β-Al_2O_3, AN
42.2	4.4	53.4	30 30	1322 1318	All glass Glass, rare β-Al_2O_3

YODER'S POINTS (UNPUBLISHED)

Weight Per Cent			Liquidus Temperature
$CaAl_2Si_2O_8$	$Ca_2Al_2SiO_7$	$NaAlSiO_4$	
Primary-Phase Nepheline			
10 10 10	33.3 28.8 24.3	56.7 61.2 65.7	1288° (checked), MEL at 1286° 1298° redetermined at 1302° 1314°
Primary-Phase Carnegieite			
10 10	21.6 18.9	68.4 71.1	1331° 1348°

therms in the two fields, is not very pronounced; it is but 3° higher than the temperature of the pseudoeutectic on the limiting $NaAlSiO_4$–$Ca_2Al_2SiO_7$ join. In a truly ternary system there is a maximum temperature on the boundary curve at the point where the line joining the composition of the two solid phases in equilibrium with a liquid intersects the boundary curve. This fact can sometimes be used as an indication of the composition of the solid phases, but no such interpretation can be made in the present case. The compositions of the phases do not lie in the plane of the system, and the solid solutions cannot be treated in ternary fashion. It does not necessarily follow that a composition on the gehlenite-nepheline side of the maximum will completely crystallize as some point "below" the maximum, i.e., in a direction away from anorthite, as it would if the maximum were present in the boundary curve of a ternary system.

The junction of the fields of plagioclase, melilite, and nepheline solid solutions is at a composition approximately 39½ per cent anorthite, 27½ per cent gehlenite, and 33 per cent nepheline and at a temperature of 1,266° C. This point is not a ternary eutectic, as crystallization is not completed at the temperature indicated (1,266°), nor does this point represent the composition of the last residual liquid. It is merely the intersection of the plane of the system with the univariant boundary curve between the quaternary fields of melilite, plagioclase, and nepheline solid solutions.

The boundary curves are not curves of univariant equilibrium but are divariant; the change in composition of a liquid in the system that is undergoing crystallization is not along a boundary curve, as it is in a ternary system. The nepheline solid solution–carnegieite solid solution boundary curve is not an isotherm, as the two extremities are not at the same temperature. Inversion from the high- to the low-temperature form begins at 1,352° C. at the $CaAl_2Si_2O_8$–$NaAlSiO_4$ join, and at 1,297° C. at the $Ca_2Al_2SiO_7$–$NaAlSiO_4$ join, with intermediate temperatures along the boundary between the two.

THE PLANE $CaAl_2Si_2O_8$–$Ca_2Al_2SiO_7$–$NaAlSiO_4$ PLUS 10 PER CENT $CaSiO_3$

To obtain more definite information on courses of crystallization, and, by so doing, to obtain some information as to the composition of the mix-crystals (with particular reference to the melilites) that form from melts in this system, it was deemed desirable to trace the course of the boundary curve between the fields of plagioclase, melilite, and nepheline solid solutions within the quaternary system. As has been seen, the pseudoeutectic point marks the position and temperature of this curve as it passes through the plane of the system here studied.

The technique used was to locate the pseudoeutectic (the point where plagioclase, melilite, and nepheline solid solutions crystallize simultaneously) in some other plane of the tetrahedron. The plane of $CaAl_2Si_2O_8$–$Ca_2Al_2SiO_7$–$NaAlSiO_4$ plus 10 per cent $CaSiO_3$ within the subsidiary tetrahedron anorthite-gehlenite-nepheline-wollastonite was chosen as a convenient one. Figure 8 represents the subsidiary tetrahedron, oriented as in Figure 1 but removed from the parent-tetrahedron and enlarged for clarity. The dotted lines indicate the plane of 10 per cent wollastonite. The location of the pseudoeutectic was determined in this plane with but three quenching runs, and the third composition that was made up fortunately corresponded exactly with the desired point on the

boundary curve. These data are given in Table 2.

There have now been located two points on the univariant boundary curve in the quaternary system between the fields of plagioclase, melilite, and nepheline solid solutions. The two four-phase points have the following parameters:

1. Composition = 39½ per cent anorthite, 27½ per cent gehlenite, 33 per cent nepheline; temperature = 1266° C.
2. Composition = 35.7 per cent anorthite, 17.8 per cent gehlenite, 36.5 per cent nepheline, 10 per cent wollastonite; temperature = 1248° C.

If this curve were to be accurately determined, numerous points within the quaternary system would have to be located. It is here assumed that a straight line joining the two determined points represents the boundary curve. If this curve is to be used as a reference line, the temperature gradient along it must also be known; and again the assumption is made that a linear relationship holds, defined by the temperatures of the two measured points. That these assumptions are not entirely correct can be shown by consideration of the system $CaSiO_3$–$CaAl_2Si_2O_8$–$NaAlSiO_4$.[29] This system, which forms one face of the subsidiary tetrahedron as shown in Figure 8, is encroached upon by the field of melilite. Therefore, at the junction of the fields of melilite, plagioclase, and nepheline solid solutions, there is located a third point on the quaternary boundary curve. Gummer's diagram[30] shows this point to be approximately 34 per cent anorthite, 38 per cent nepheline, and 28 per cent wollastonite, and at a temperature of 1,190° C. Consideration of the three points shows that the temperature drop along the boundary is not linear, and a projection of the compositions of the three indicates that the boundary is not straight. The departure from linearity between the values of 0–28 per cent $CaSiO_3$ is not considerable, however, and errors introduced by assuming linear values between 0 and 10 per cent $CaSiO_3$ are small enough to justify the assumptions. The complications introduced by solid solutions bring up the possibility of a temperature maximum along the boundary curve, in which case any assumptions as to temperature gradient are, of course, nullified. It is not probable that such a situation exists, however, in the limited extent of the boundary under consideration. Thus knowledge of the temperature at which nepheline solid solution and plagioclase simultaneously crystallize fixes the composition of the liquid at a point on the boundary curve corresponding to this temperature.

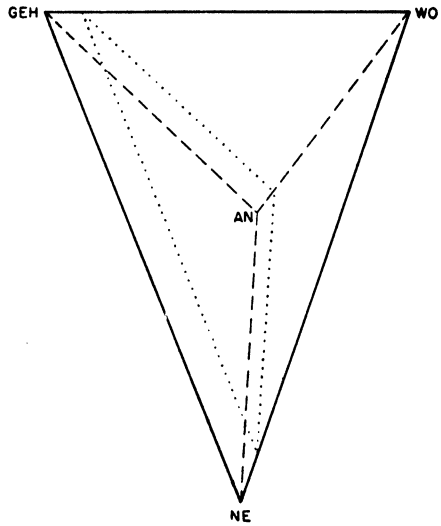

FIG. 8.—The subsidiary tetrahedron AN-GEH-NE-WO, showing (*dotted lines*) plane of 10 per cent WO.

COURSES OF CRYSTALLIZATION

The fact that the composition of the crystalline phases does not lie in the plane

[29] See ftn. 21.

[30] P. 515 of ftn. 21.

of the system and, consequently, that the residual liquids resulting from crystallization also leave this plane makes difficult the tracing of courses of crystallization. If the compositions of the solid solutions were precisely known, the courses of crystallization viewed in three dimensions within the quaternary system could be precisely described. The present state of uncertainty as to the composition of these solid solutions, with the possible exception of the plagioclase, makes this impossible.

The plagioclase-melilite-nepheline boundary curve was located for the purpose of obtaining data on the courses of crystallization and has been used in the present work to locate what may be called a "four-phase boundary" within the quaternary field of melilite. The technique of locating three-phase boundaries in a ternary system has been described by N. L. Bowen.[31] In the present investigation the method used was to locate a composition in the melilite field from which plagioclase and nepheline solid solutions simultaneously crystallized as secondary phases. The significance of the simultaneous crystallization of the secondary phases is that the crystallization of the primary phase, melilite, results in a liquid of changing composition (increasingly deficient in the melilite composition), the composition path of which directly hits the *melilite-plagioclase-nepheline boundary curve* in the quaternary system. Thus the location of such a point in the field of melilite, coupled with the observation of the temperature at which the plagioclase and nepheline solid solution crystal-

TABLE 2

THERMAL DATA FOR THE PLANE $CaAl_2Si_2O_8$–$Ca_2Al_2SiO_7$–$NaAlSiO_4$ + 10 PER CENT $CaSiO_3$

Weight Per Cent				Time (Minutes)	Temperature (° C.)	Final Condition
$CaAl_2Si_2O_8$	$Ca_2Al_2SiO_7$	$NaAlSiO_4$	$CaSiO_3$			
Primary-Phase Melilite						
$37\frac{2}{3}$	$21\frac{2}{3}$	$30\frac{2}{3}$	10	30	1275	All glass
				30	1271	Glass, MEL
				30	1263	Glass, MEL, AN
Primary-Phase Anorthite						
37.2	19.7	33.1	10	30	1267	All glass
				30	1263	Glass, AN, very rare MEL
Simultaneous Phases Anorthite-Melilite-Nepheline						
35.7	17.8	36.5	10	30	1250	All glass
				30	1246	Glass, MEL, NE, sparse AN

[31] "The Crystallization of Haplobasaltic, Haplodioritic, and Related Magmas," *Amer. Jour. Sci.*, Vol. XL (4th ser., 1915), pp. 171–74.

lize, fixes two points on a crystallization path in the melilite field. As described above, the boundary curve has been located within the quaternary system, and the temperature of appearance[32] of the secondary phases locates the point on the boundary curve that is "hit" by the crystallizing liquid. A line joining these two points must pass through the composition of the crystals that have separated from the liquid. The above considerations are applicable in principle to the field of nepheline, although determinations in this field have not been carried out.

Examination of Figure 6 shows three points near the pseudoeutectic in the field of melilite. These points, along with others to be subsequently discussed, are reproduced in Figure 9, which is an enlarged section of the equilibrium diagram.

If the system were ternary, the final temperature of consolidation of these three compositions would be 1,266°, and at the composition of the eutectic. Runs made on point 1 (Fig. 9) did not show nepheline until approximately 1,260° C., and plagioclase appeared at approximately 1,245° C. In point 2, plagioclase was present at approximately 1,260° C., and nepheline appeared at 1,255° C. Point 3 represents a composition in which plagioclase and nepheline solid solutions appear simultaneously and at a temperature of 1,257° C. Thus the quaternary boundary curve is encountered at a temperature of 1,257° C. for this particular initial composition. It has been shown that this boundary curve passes through the plane of 10 per cent wollastonite at 1,248° C. Assuming a linear temperature relationship, 1,257° C. is halfway between the two reference temperatures of 1,248° and 1,266° C.; and thus the boundary curve is encountered at approximately the plane of 5 per cent wollastonite within the subsidiary tetrahedron *AN-GEH-NE-WO*. This point will be more fully discussed in the following section, but it can be seen from examination of Figure 1 that this course of crystallization is roughly toward the SiO_2 corner of the main tetrahedron.

If the nepheline solid solutions were ternary, point *4* of Figure 9 would also

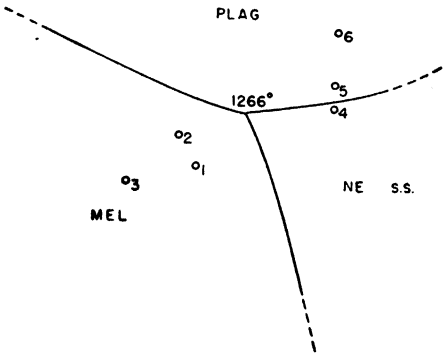

FIG. 9.—Enlarged portion of the equilibrium diagram for the system $CaAl_2Si_2O_8$–$Ca_2Al_2SiO_7$–$NaAlSiO_4$.

consolidate at 1,266° C. Plagioclase appears in runs on this point at 1,267° C., and melilite does not appear until the temperature has fallen to 1,248° C. Point *5*, in the plagioclase field, precipitates nepheline at 1,273° C. and melilite at approximately 1,246° C., whereas in point *6* melilite appears at approximately 1,255° C. The fact that melilite does not appear in runs made on points *5* and *6* until temperatures well below that of the pseudoeutectic are reached is indicative of the rather strongly nonternary nature of the nepheline solid solutions. Even in compositions close to the pseudoeutectic, there is a considerable difference in tem-

[32] If devitrified mixtures are used, this temperature is that of the simultaneous disappearance of the secondary phases.

perature between that of the pseudo-eutectic and that of appearance of the third crystalline phase, indicating that the composition of the liquid is well out of the plane of the system.

No attempt has been made to follow courses of crystallization within the fields of β–Al_2O_3 or carnegieite. In general, compositions within the field of β–Al_2O_3 will follow a path away from the Al_2O_3 corner of the tetrahedron.

THE SUBSTANCES IN SOLID SOLUTION

With the exception of quartz, all rock-forming minerals are members of solid-solution series. Pure end-members of solid-solution series are rarely, if ever, found in nature; and it has been amply demonstrated that the pure end-members are never formed in synthetic melts when there are substances present with which they may form solid solutions. The nature of the substances in solid solution in the nepheline and carnegieite of this system is unsettled, but certain data here found tend to limit the possibilities at least with respect to nepheline. The "four-phase boundary" in the melilite field has, moreover, brought to light the probable existence of a hitherto unknown "molecule" in solid solution with gehlenite.

ANORTHITE

It was earlier thought that relatively pure anorthite could form in melts containing nepheline, but it is now known that the pure end-members of the plagioclase series cannot form in melts containing both soda and lime. Although few optical data could be obtained on the feldspars because of their small size and platy habit, enough were obtained to indicate that an anorthite-rich plagioclase rather than pure anorthite is formed. Crystals formed in a mix with but little soda (approximately 6 per cent nepheline, 62 per cent anorthite, 32 per cent gehlenite) had a maximum index that very closely matched the glass at a value of $n = 1.588$, indicating that virtually pure anorthite is here formed. At a somewhat higher soda content (16 per cent nepheline, 46 per cent anorthite, 38 per cent gehlenite), the maximum plagioclase index again matched the index of the glass, at a value of 1.586. In two mixes near the plagioclase-nepheline boundary curve maximum values of $n = 1.585$ were observed, the nepheline content here reaching 37 per cent. The value of 1.585 for the lowest observed maximum plagioclase index indicates that the anorthite content probably does not drop below 95–96 per cent in this system.

Extinction angles of 72° and 74° on two well-formed crystals were observed, as measured from the albite twinning lamellae to α. These crystals were those formed from melts with the 37 per cent nepheline content. Reference to standard tables indicates an anorthite content of approximately 98 per cent, as against 95–96 per cent on the basis of index of refraction. No discrepancy exists, however, as recent work has shown that plagioclases formed at high temperatures do not have the same composition–extinction-angle relation as do the low-temperature plagioclases, for which standard tables have been compiled.[33, 34] Plagioclases formed at high temperatures show an apparently greater anorthite content on the basis of extinction angle, as is here evident.

[33] A. Köhler, "Die Abhängigkeit der Plagioklasoptik vom vergangenen Warmeverhalten," *Min. u. petrog. Mitt.*, Vol. LIII (1942), pp. 159–79.

[34] See bibliography in paper by Christoffer Oftedahl, "High Temperature Optics in Plagioclases of the Oslo Region," *Norsk. geol. Tids.*, Vol. XXIV (1944), pp. 75–78.

β-ALUMINA

The probable composition of β–Al_2O_3 has been discussed.

NEPHELINE

Several substances are known to enter into solid solution with nepheline in considerable amount. A complete series of solid solutions between $KalSiO_4$ and $NaAlSiO_4$ exists,[35] but potash compounds are excluded in this study. Experimental work has shown that 35 per cent each of albite[36] and anorthite[37] are taken into solid solution. The plagioclases are sufficient to account for the excess silica, as well as the calcium that is found in natural nephelines; but A. N. Winchell[38] has suggested that the molecule $CaAlAlO_4$ may also play a part in the composition.

The high indices found by Smalley in his work on the system $NaAlSiO_4$–$Ca_2Al_2SiO_7$[39] are considerably above those brought about by solid solution of anorthite in nepheline. The inversion relations of nepheline and carnegieite were also such that anorthite could not be considered as the material entirely responsible. In the present study no negative nephelines were observed; those formed from melts closest to the field of carnegieite (at approximately 60 per cent nepheline) were sensibly isotropic, those precipitated from all other melts poorer in the nepheline molecule were positive, becoming more strongly birefringent as the $NaAlSiO_4$ content of the liquid decreased. The nepheline indices, as in Smalley's system, increase with increasing $Ca_2Al_2SiO_7$ content of the melt. Isotropic crystals with $n = 1.540$ were observed in the mixture at 5 per cent $Ca_2Al_2SiO_7$, 60 per cent $NaAlSiO_4$, 35 per cent $CaAl_2Si_3O_8$; and values of $\omega = 1.548$, $\epsilon > 1.548$, were observed in crystals near the plagioclase-nepheline boundary curve at a value of roughly 20 per cent $Ca_2Al_2SiO_7$.

It was noted that, in mixtures from which nepheline crystallized as a primary or secondary phase and melilite as the tertiary phase, the temperature of appearance of the melilite was considerably below that of the pseudoeutectic. The interpretation was that the quaternary boundary curve between the fields of melilite, plagioclase, and nepheline solid solutions was reached in the direction of falling temperature. That is, the temperature along this boundary falls in the direction of the direction of the anorthite-nepheline-wollastonite face of the subsidiary tetrahedron, and the composition of the liquid must change in this direction upon crystallization of nepheline solid solutions. Thus plagioclase cannot be the sole substance in solid solution in the nephelines of this system. In addition, anorthite or any substance near anorthite in composition would not account for the strong nonternary tendency that is shown in the nepheline solid solutions; the plagioclase compositions lie nearly in the same plane as that of this system; anorthite, of course, lies in the plane.

It thus appears that at least part of the material entering into solid solution with nepheline lies on that side of the system that is toward the Al_2O_3 corner of the main tetrahedron. Sodium or calcium aluminates would fulfil this requirement. The $CaAlAlO_4$ molecule proposed by Winchell might well be a substance en-

[35] N. L. Bowen, "The Sodium-Potassium Nephelites," *Amer. Jour. Sci.*, Vol. XLIII (4th ser., 1917), pp. 115–32.

[36] P. 106 of ftn. 7.

[37] P. 563 of ftn. 4.

[38] *Elements of Optical Mineralogy*, Part II (3d ed.; New York: John Wiley & Sons, Inc., 1933), p. 297.

[39] See ftn. 10.

tering into solid solution with nepheline in this system. It is also known that $Na_2Al_2O_4$ forms a solid-solution series with carnegieite,[40] and it is not unlikely that it can also form solid solutions with nepheline. The establishment of "four-phase boundaries" in the field of nepheline might shed more light on the exact nature of the solid solutions.

At first sight it appears peculiar that no evidence of $CaAlAlO_4$ and/or $Na_2Al_2O_4$ is found in naturally occurring nephelines. An excess of SiO_2 is usually present, and the above compounds would make for undersilicated nepheline. The system studied is deficient in silica, however, as compared to natural magmas, and it is likely that in magmas the silica content is always sufficient so that feldspar forms solid solutions with nepheline rather than the aluminates. It is more probable, however, that at the high temperatures of this work, probably higher than those of magmas, solid solutions form that are not found in nature.

CARNEGIEITE

This investigation has produced no new evidence as to the nature of the carnegieite solid solutions. Smalley[41] has observed a marked index rise and inversion lowering in the system $NaAlSiO_4$–$Ca_2Al_2SiO_7$. Inversion lowering is very unusual in silicates, as it requires a greater amount of solid solution in the high-temperature form than in the low-temperature form of an enantiotropic pair. However, as the carnegieite solid solutions undoubtedly present a problem very similar to that of the nepheline mixcrystals, the situation is perhaps not too obscure. Here anorthite is unquestionably out of consideration as the responsible agent, as it is taken up by carnegieite only to the extent of 5 per cent, whereas nepheline will take 35 per cent.[42] C. E. Tilley[43] found that Na_2SiO_3 entered into carnegieite solid solutions to the extent of 24 per cent but not at all into nepheline. The changes in birefringence and index rise were not consistent with those found by Smalley, however, and it is unlikely that Na_2SiO_3 plays a part in this system. Smalley found that the trend of the solid solutions was to increase in concentration with increase in the $Ca_2Al_2SiO_7$ content of the mixture and suggested that the sodium aluminate ($Na_2Al_2O_4$) was responsible; the decrease in silica brought about by the addition of gehlenite would account for the trend rather than the addition of lime, with formation of a lime-bearing solid solution. This view is in line with the discussion of the nepheline solid solutions, and, in addition, it may be that the $CaAlAlO_4$ molecule is also present in the mixcrystals.

GEHLENITE

The indices of the melilites that crystallize in this system are significantly lower than those of pure gehlenite. Gehlenite has indices of $\omega = 1.669$ and $\epsilon = 1.658$. The melilite formed from a mixture of composition anorthite 16 per cent, gehlenite 67 per cent, and nepheline 17 per cent had indices of $\omega = 1.665$, $\epsilon' = 1.656$; and in a composition of anorthite 38 per cent, gehlenite 32 per cent, and nepheline 30 per cent the indices were $\omega = 1.656$, $\epsilon' = 1.645$; at anorthite 5 per cent, gehlenite 37 per cent, and nepheline 58 per cent, $\omega = 1.651$, $\epsilon' = 1.641$. A progressive decrease

[40] Unpublished work of J. F. Schairer and N. L. Bowen.

[41] See ftn. 10.

[42] P. 563 of ftn. 4.

[43] "The Ternary System Na_2SiO_3–$Na_2Si_2O_5$–$NaAlSiO_4$" *Min. u. petrog. Mitt.*, Vol. XLIII (1933), p. 414.

in index with increasing soda content is thus observed, a decrease that was also found by Smalley to exist in the limiting $Ca_2Al_2SiO_7$–$NaAlSiO_4$ system. No decrease in birefringence was observed.

The natural melilites have complex solid-solution relations, and this complexity has elicited considerable discussion. A complete series of solid solutions exists between $Ca_2Al_2SiO_7$ and $Ca_2MgSi_2O_7$ (akermanite); but for purposes of this discussion the magnesia-bearing varieties, as well as the iron melilites, will be neglected. In 1924, A. N. Winchell[44] took up the melilite problem from the viewpoint of volume isomorphism and largely discredited the sarcolite molecules which had been proposed earlier by W. T. Schaller[45] and partially verified by A. F. Buddington.[46] Winchell pointed out the essential R_5O_7 character of the melilites and proposed the molecules $Na_2Si_3O_7$ and $Ca_3Si_2O_7$ as theoretically capable of entering the gehlenite structure. In addition, the possibility of the presence of several sodium-aluminum silicates and a calcium-aluminum silicate was considered. Berman[47] in 1929 reviewed the problem and modified Winchell's fundamental R_5O_7 to $X_2Y_3O_7$. He also changed Winchell's $Ca_3Si_2O_7$ to $CaSi_3O_7$. The melilite molecules, as seen by Berman (omitting akermanite) are shown in the accompanying tabulation.

Gehlenite	$Ca_2Al_2SiO_7$
Soda melilite	$Na_2Si_3O_7$
Submelilite	$CaSi_3O_7$

[44] "The Composition of Mellilite," *Amer. Jour. Sci.*, Vol. VIII (5th ser., 1922), pp. 375–87.

[45] "Mineralogic Notes," Ser. 3, *Bull. U.S. Geol. Surv.*, *No. 610* (1916), pp. 109–28.

[46] "On Some Natural and Synthetic Melilites," *Amer. Jour. Sci.*, Vol. III (January, 1922), pp. 35–87.

[47] Harry Berman, "Composition of the Melilite Group," *Amer. Min.*, Vol. XIV, No. 11 (November, 1929), pp. 389–407.

On the basis of chemical analyses Berman stated that the "soda melilite" does not exceed 25 per cent in natural occurrences and the "submelilite" not over 10 per cent.

The location of the quaternary boundary curve between the fields of melilite, plagioclase, and nepheline solid solutions and the determination of the "four-phase boundary" discussed in the section on courses of crystallization was made principally to obtain information on the nature of the melilite mix-crystals in this system.

The essential data with respect to the "four-phase boundary" are here summarized. A liquid represented by the composition of point *3* in Figure 9, upon crystallizing melilite, hits the quaternary boundary curve at 1,257° C., as evidenced by simultaneous crystallization of plagioclase and nepheline as secondary phases. Two points are thus located; the initial composition (point *3*, Fig. 9), and that point on the boundary curve representing the composition of the liquid at the time when sufficient melilite has crystallized to attain this composition. A straight line, drawn from this point on the boundary curve through the initial composition, points to the composition of the melilite crystallizing at the time that the boundary curve is reached. The composition of the initial point is known, and that of the point on the boundary curve can be located by making the assumptions of a straight boundary curve with linear temperature drop, as earlier discussed. As the temperature is 1,257°, half of the difference between the planes of 0 and 10 per cent wollastonite can be assumed to be in the plane of 5 per cent wollastonite. Calculation of this point gives the following oxide percentages,

and listed with it is the initial composition of the mixture:

		Per Cent
Initial composition	CaO	21.92
	Na_2O	6.26
	Al_2O_3	36.62
	SiO_2	35.20
Point on boundary curve	CaO	19.11
	Na_2O	7.67
	Al_2O_3	34.64
	SiO_2	38.55

Figure 10 represents the subsidiary tetrahedron, as in Figure 8. The quaternary boundary curve is diagrammatically indicated by the slightly curved line. Points *a* and *b* represent, respectively, the initial composition (point *3*, Fig. 9) and the point on the boundary curve hit by this point upon crystallization. The arrow, joining the two, points to the composition of the melilite forming when point *b* is reached.

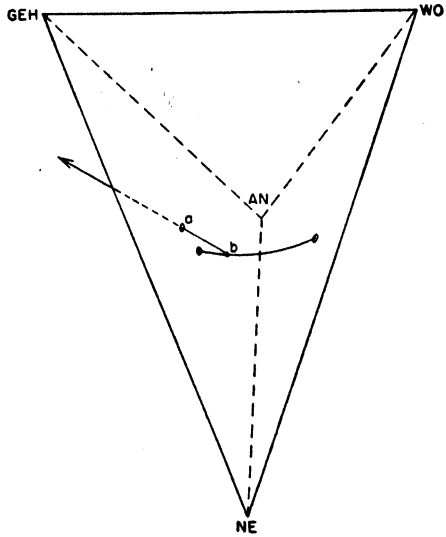

FIG. 10.—The subsidiary tetrahedron AN-GEH-NE-WO, showing quaternary boundary curve and four-phase boundary.

It is evident that the melilite solid solution crystallizing from the composition of point *a* is represented by some point "below" the plane of this system, i.e., on the side toward the Al_2O_3 corner of the main tetrahedron. Thus none of the hypothetical molecules proposed by Winchell or Berman enter into solid solution in this system. Considering pure gehlenite, $Ca_2Al_2SiO_7$, the following replacements can be made:

$CaCaAlAlSiO_7$ (gehlenite)
$CaAlAlAlAlO_7$ (Al for Ca and Al for Si)

This molecule, conforming to the gehlenite structure, can be written $CaO \cdot 2Al_2O_3$ and will be referred to as "1:2." The 1:2 "molecule" is not known to exist as an independent compound,[48] but this fact does not preclude its entering into solid solution as a component. If it does form a solid solution with gehlenite, the composition of the mix-crystals would be represented by a line joining the two components within the principal tetrahedron. The extension of the join between points *a* and *b* (Fig. 10) should then meet this line at some point representing the composition of the mix-crystals formed from a melt of composition *a*.

An orthographic projection of the lines representing the joins between *a* and *b* and gehlenite–1:2 shows that they come very close to intersecting.[49] This fact could also be proved by analytical means, either by obtaining the equations of the two lines or by showing that all four points lie in a plane.

[48] P. 306 of ftn. 2.

[49] A precise join would be fortuitous, because of experimental errors, here listed:

1. The primary phase of point *3*, Fig. 9, is melilite, the high relief of which tends to mask the appearance of the secondary phases. The determination of simultaneous phases, as well as the precise temperature of appearance, is difficult.

2. Points *a* and *b* are relatively close together in terms of the main tetrahedron; thus a slight error in the join between them is magnified in the extension of the join. If a point such as *a* were located at a greater distance from the boundary curve, the

On the basis of this single determination, it thus appears that the 1:2 compound of lime-alumina is the substance in solid solution in gehlenite crystallizing from this system. Moreover, if any other hypothetical "molecule" enters the mix-crystals, it must do so in very small amounts.

As was postulated for the nepheline solid solution, it is again a subsilicious phase that precipitates from melts of the present investigation. No such silica-poor melilites are found in nature. Natural magmas are not comparable in composition to melts in this system, however; and, as previously discussed, the temperatures of crystallization are considerably lower in magmas. This difference of composition of the natural magmas presumably results in the formation of solid solutions of a nature other than those containing the 1:2 molecule, and the understanding of the complexity of the melilites is by no means completely resolved by this investigation.[50]

PETROLOGIC SIGNIFICANCE

Complete knowledge of the equilibrium relations within the soda-lime-alumina-silica system is of great importance in the understanding of the problems of igneous petrology, and the present paper is a contribution toward that end. These four oxides constitute, by weight, 90 per cent of the average of 546 granites, 85 per cent of the average of 43 nepheline syenites, and 77 per cent of the average of 161 basalts.[51] The minerals—quartz, plagioclase feldspars, nepheline, wollastonite, gehlenite, corundum, sillimanite, mullite, grossularite, larnite, and jadeite—are formed of these four oxides, although the last five are formed only under "metamorphic conditions." Knowledge of the four oxide systems is also of prime importance in certain technological fields.

The equilibrium relations as determined in this study contribute to the knowledge of the relations within the main system. All seven faces of the subsidiary double tetrahedron have now been investigated.[52, 53] It has been mentioned that the melts within the plane here studied do not correspond in composition to any natural magmas. The closest approach is to be found in magmas that formed certain basic alkaline rocks, such as the tephrites, melilite-nephelinites, urtites and ijolites, turjaites, and melilite basalts. All these rocks contain greater or lesser amounts of a pyroxene and numerous accessory

identification of secondary phases (1, above) would be more difficult.

3. The assumption of a straight boundary curve with linear temperature drop is not strictly true.

4. There are experimental errors of temperature measurement and composition synthesis.

[50] Although the results of this investigation as interpreted above indicate solid solution of $CaO \cdot 2Al_2O_3$ in gehlenite, certain difficulties are encountered in this supposition. No solid solutions of this nature were encountered in the investigation of the system CaO–Al_2O_3–SiO_2 (ftn. 26). The proposed substitution of Al for Ca in the solid solutions is rather doubtful, owing to the small size of the Al atom. It is therefore important to note that another interpretation is possible: *if* a maximum temperature is present on the melilite-nepheline-plagioclase boundary curve, the melilites crystalizing from this system may be *enriched* in Si and Na. Thus, if a temperature of 1,257° exists on the low-silica side of the plane of the system, the "molecule" in solid solution in gehlenite could be the $Na_2Si_3O_7$ proposed by Berman, which, it must be admitted, is more likely from a structural point of view. It is hoped that this point can be cleared up in a future investigation.

[51] R. A. Daly, *Igneous Rocks and the Depths of the Earth* (New York: McGraw-Hill Book Co., Inc., 1933, pp. 9–28.

[52] See bibliography in paper by N. L. Bowen, "Phase Equilibria Bearing on the Origin and Differentiation of Alkaline Rocks," *Amer. Jour. Sci.*, Vol. CCXLIII*A* (Daly Vol.) (1945), p. 83.

[53] The anorthite-gehlenite-wollastonite face is based on unpublished work of V. C. Juan.

minerals. The melilite of this system is also less complex than the natural mineral, because of the absence of magnesia and iron.

The strong tendency of gehlenite, nepheline, and anorthite to form solid solutions is again demonstrated in the present study; and, because of the nonternary nature of the solid solutions, the system must be considered in terms of the four constituent oxides. Crystallization of the nepheline and melilite solid solutions results in a liquid enriched in SiO_2 and Na_2O; and, once the quaternary boundary curve between the fields of melilite, plagioclase, and nepheline is reached, fractional crystallization could carry the residual liquid toward the plane of anorthite-nepheline-wollastonite, which is common to the subsidiary double tetrahedron. The fact that the field of melilite plunges through this boundary face, as discussed by Bowen,[54] results in a quaternary reaction point within the adjoining tetrahedron, anorthite-wollastonite-nepheline-albite, and permits fractionation to continue so that residual liquids within this portion of the main tetrahedron can result. Bowen[55] points out that fractionation can continue until the liquid reaches a composition within that portion of the quaternary system outlined by the compounds nepheline-albite-devitrite ($Na_2O \cdot 3CaO \cdot 6SiO_6$)–sodium disilicate, the crystalline assemblages running the gamut from simplified melilite-nephelinites through simplified tephrites and phonolites to alkali rhyolites, with a soda-rich liquor remaining.

The encroachment of a field of alumina upon the anorthite-nepheline join is suggestive in connection with the occurrence of corundum in certain anorthosites and nepheline-bearing syenites.

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[54] See reference cited in ftn. 52.

[55] See reference cited in ftn. 52.

THE COMPOSITION OF METEORITIC MATTER

I. THE COMPOSITION OF THE SILICATE PHASE OF STONY METEORITES

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ABSTRACT

A statistical study has been made of the analyses of the silicate phases of 107 selected stony meteorites. Frequency-distribution curves have been plotted, and arithmetic means, together with standard deviations and precisions, have been calculated for the major constituents.

INTRODUCTION

At the present time, meteorites constitute one of the more important sources of information concerning the composition of the solar system. A thorough study of the composition of meteoritic matter can, conceivably, give valuable clues for the solution of such diverse problems as the geochemical distribution of the elements, the process of planet formation, and the origin of the elements.

In 1938, V. M. Goldschmidt,[1] utilizing primarily chemical data obtained on meteorites, compiled a table of "cosmic abundances" of nuclear species, which is of particular interest to nuclear physicists and astrophysicists because of its bearing upon the problems of the origin of the elements and the relationships which may exist between "cosmic abundance" and nuclear properties. Since that time, interest in the problem of the origin of the elements has progressed to a point at which further inquiry into the experimental and theoretical foundations for such a compilation is desirable. This paper represents the first phase of such an inquiry.

As our first step, we have attempted to answer two questions: (1) How accurately can the average composition of the silicate phase of stony meteorites be expressed at the present time? (2) For a given constituent, how many meteorites must be analyzed if the probable error of the arithmetic mean of the concentration is to lie below a certain fixed value? We have attempted to answer these questions by analyzing statistically the available data on the major constituents of stony meteorites.

THE PRESENT INQUIRY

In 1916, G. P. Merrill[2] averaged the composition of 59 stone meteorites, most of them known falls. Goldschmidt,[3] in his compilation of relative abundances, used Merrill's averages, after first subtracting the iron-nickel and sulphide phases, in that manner obtaining the average composition of the silicate phase. Since Merrill's averages were obtained, a sufficient number of additional meteorite falls have been analyzed to permit a statistical study of the distribution of the elements in the silicate phase.

The individual analyses for 107 selected stony meteorites are given in Table 1. Of the analyses, 58 are from

[1] "Geochemische Verteilungsgesetze der Elemente. IX. Die Mengenverhältnisse der Elemente und der Atom-Arten," *Skrifter utgitt av Det Norske Videnskaps-Akademi i Oslo. I Mat.-Naturv. Klasse, 1937*, No. 4 (1938).

[2] "Report on Researches on the Chemical and Mineralogical Composition of Meteorites, with Especial Reference to Their Minor Constituents," *Mem. Nat. Acad. Sci.*, Vol. XIV (1916).

[3] See ftn. 1.

TABLE 1

TABLE OF SELECTED ANALYSES OF STONY METEORITES

Meteorite*	Silicate Phase																Iron-Nickel Phase					Ref.
	SiO_2	MgO	Fe†	Al_2O_3	CaO	Na_2O	K_2O	Cr_2O_3	NiO	MnO	CoO	TiO_2	H_2O	P‡	O§	Total	Fe	Ni	Co	S	Total	
Aguila Blanca, Argentina	48.19	27.79	9.86	3.95	2.11	1.71	0.26	0.75	2.34	0.22	Tr.	Tr.	.....	0.25	2.88	100.39	15.58	1.98	0.03	1.06	18.65	1
Alfianello, Italy	41.58	25.89	21.15	1.97	0.98	1.20	0.27	0.46	.....	0.14	.....	.....	.....	0.17	6.06	99.87	5.74	1.14	0.08	2.54	9.50	2
Allegan, Mich.	48.57	30.56	9.15	3.54	2.40	0.92	0.32	0.74	Tr.	0.25	.....	0.11	0.35	0.29	2.71	99.91	23.31	1.81	0.15	1.84	28.11	2
Angra dos Reis, Brazil	44.51	10.18	6.74	8.84	24.83	0.26	0.19	Tr.	.....	.....	.....	2.42	.....	0.10	1.96	100.03	0.81	.....	.....	0.45	1.26	2
Baroti, India	47.06	29.31	13.25	2.85	2.12	1.08	0.05	0.21	.....	Tr.	.....	0.19	0.20	0.23	3.87	100.42	12.45	0.75	0.03	2.47	15.70	2
Beaver Creek, Canada	48.28	30.61	10.75	2.80	2.27	1.02	0.13	0.27	0.04	0.30	.....	0.09	0.36	0.24	3.16	100.33	18.74	1.50	0.09	1.84	22.17	2
Bettrechies, France	47.71	27.96	16.79	0.22	2.56	0.64	0.22	.....	0.25	0.15	0.02	0.02	.....	0.03	4.95	101.52	14.64	0.71	0.07	3.87	19.29	3
Bishopville, S.C.	57.32	33.68	1.22	1.72	2.03	1.04	0.09	.....	0.54	0.19	Tr.	.....	2.00	.....	0.35	100.18	0.18	0.30	.....	0.04	0.52	2
Bjurbole, Finland	46.97	29.46	12.27	2.92	2.08	1.42	0.37	0.67	0.08	0.14	.....	.....	.....	0.16	3.52	100.06	9.84	0.72	0.04	1.98	12.58	2
Bluff, Tex.	39.89	27.44	19.59	2.30	2.33	.....	.....	.....	1.68	0.48	0.17	.....	.....	0.20	5.67	99.75	3.47	0.65	0.09	1.30	5.51	2
Bowden, S. Africa	47.55	30.10	10.25	3.77	2.89	0.99	0.14	0.42	.....	0.31	.....	0.16	0.35	.....	3.10	100.03	19.86	1.49	0.06	2.15	23.56	4
Cape Girardeau, Mo.	46.53	31.06	11.45	2.97	1.81	1.12	0.14	0.60	.....	.....	.....	.....	0.61	0.29	3.37	99.95	20.07	1.32	0.11	2.07	23.57	2
Colby, Wis.	45.28	32.16	12.81	3.10	—	1.21	0.15	0.55	0.23	0.49	0.02	.....	.....	0.20	3.74	99.94	5.73	0.07	0.02	2.76	8.58	5
Collescipoli, Italy	42.51	26.02	15.16	7.90	2.20	1.79	0.32	.....	.....	.....	.....	.....	.....	.....	4.34	100.24	17.04	1.49	0.09	.....	18.62	2
Cranstad, S. Africa	48.78	30.93	9.34	3.16	1.90	0.88	0.09	0.48	0.13	0.05	.....	0.39	0.45	0.25	2.75	99.58	20.00	1.55	0.23	1.88	23.66	6
Crumlin, Ireland	47.49	29.57	12.00	2.49	2.38	1.25	0.13	0.70	.....	0.31	.....	0.20	.....	0.23	3.51	100.26	11.09	1.15	0.08	1.91	14.23	7
Cullison, Kan.	47.27	31.64	9.42	5.68	0.83	2.41	0.23	0.05	.....	0.39	.....	.....	.....	0.01	2.70	100.63	21.27	1.80	0.10	2.18	25.35	2
Daniels Kuil, S. Africa	58.47	32.88	0.27	2.71	2.59	0.97	0.23	0.69	0.17	0.03	.....	0.18	0.52	Tr.	0.08	99.80	28.85	1.78	0.12	3.46	34.21	6
Ekerby, Sweden	44.29	28.72	14.78	2.27	2.49	1.22	0.26	0.58	.....	0.31	.....	0.28	0.04	0.28	4.34	99.86	16.93	1.73	.....	1.90	20.56	8
Elm Creek, Kan.	36.75	25.66	20.30	3.10	1.62	—	—	0.35	.....	0.01	.....	.....	5.10	.....	5.81	98.70	5.94	0.77	0.09	.....	6.80	2
Estacado, Tex.	43.56	27.65	14.68	4.38	3.64	2.52	0.39	—	.....	—	.....	—	.....	0.18	4.20	101.20	14.68	1.60	0.08	1.37	17.73	2
Farmington, Kan.	45.74	29.95	14.04	2.05	2.00	0.84	0.13	0.66	0.37	0.18	.....	.....	.....	.....	4.02	99.98	9.86	0.94	0.06	1.82	12.68	2
Felix, Ala.	36.39	21.40	22.09	3.51	5.91	0.67	0.15	0.87	1.09	0.74	.....	.....	0.17	.....	6.33	99.32	5.62	0.36	0.08	1.73	7.79	2
Fisher, Minn.	43.69	29.38	14.40	4.79	2.19	Tr.	—	0.54	0.23	0.38	.....	.....	.....	.....	4.13	99.73	9.72	1.61	0.08	Tr.	11.41	2
Florence, Tex.	46.69	30.22	10.94	3.57	3.59	0.50	0.74	0.23	0.18	0.70	.....	Tr.	.....	0.26	3.21	100.83	19.29	1.49	0.07	1.84	22.69	9
Forest City, Iowa	47.87	32.18	10.71	2.80	1.90	1.09	0.08	0.13	0.19	.....	.....	.....	.....	.....	3.07	100.02	22.02	1.19	0.13	2.25	25.59	2
Forksville, Va.	41.30	28.40	18.80	3.25	1.46	0.53	0.28	0.10	0.17	.....	.....	.....	.....	0.19	5.44	99.92	5.66	0.78	0.04	1.14	7.62	10
Hedjaz, Arabia	44.16	32.56	10.59	4.48	3.49	0.47	0.59	0.09	—	.....	.....	0.06	—	.....	3.03	99.52	12.98	0.94	—	1.61	15.53	11
Hendersonville, N.C.	48.09	29.88	11.76	2.30	2.22	1.00	0.10	0.61	.....	.....	.....	.....	.....	0.01	3.20	99.34	2.37	0.21	0.01	1.61	4.20	2
Hessle, Sweden	45.40	30.82	12.84	1.91	2.56	1.93	.....	.....	.....	.....	.....	.....	.....	Tr.	3.68	99.14	16.36	2.16	Tr.	0.18	18.70	2
Holbrook, Ariz.	41.89	29.08	16.96	4.29	2.40	Tr.	.....	.....	0.08	0.25	.....	.....	.....	.....	4.86	99.81	9.20	0.42	0.03	2.75	12.40	2
Homestead, Iowa	47.44	30.06	11.67	2.54	2.15	1.11	0.16	0.39	.....	0.10	.....	.....	0.57	0.23	3.46	99.88	13.68	1.19	0.09	2.28	17.24	12
Hvittis, Finland	59.30	33.17	0.38	2.21	2.01	1.80	0.46	0.81	.....	.....	.....	.....	.....	0.11	0.11	100.35	24.66	1.96	0.07	3.30	29.99	2
Indarch, Russia	47.84	22.87	14.95	2.64	1.55	0.17	.....	.....	0.74	.....	0.07	.....	3.75	0.63	4.44	99.65	19.30	0.95	0.02	5.10	25.37	2
Jelica, Yugoslavia	43.21	28.03	18.31	2.08	1.36	[0.13]		0.55	.....	0.21	.....	.....	.....	.....	5.24	99.12	6.12	0.83	0.05	2.58	9.58	2
Jerome, Kan.	35.26	22.99	23.16	1.89	1.40	0.69	0.30	0.62	1.89	Tr.	Tr.	Tr.	3.23	0.29	6.73	98.45	3.81	0.43	0.01	1.88	6.13	2
Johnstown, Colo.	50.76	27.12	11.40	2.37	2.65	0.33	0.10	1.03	.....	0.40	.....	.....	.....	0.02	3.27	99.45	0.55	.....	.....	0.35	0.90	13
Juvinas, France	48.04	7.21	14.66	13.51	10.61	0.55	—	Tr.	0.11	.....	.....	0.57	.....	—	3.99	99.46	Tr.	.....	Tr.	0.06	0.06	2
Karoonda, Australia	35.54	25.99	21.94	5.81	2.70	0.74	0.27	0.51	.....	0.22	.....	—	0.14	0.20	6.55	100.81	2.95	0.02	.....	1.45	4.42	14
Kernouve, France	44.48	31.97	12.27	4.31	2.55	1.90	Tr.	Tr.	.....	.....	.....	.....	.....	.....	3.52	101.00	22.25	1.55	.....	2.15	25.95	2
Kesen, Japan	45.44	28.58	16.21	Tr.	2.78	0.56	0.76	.....	.....	0.34	.....	.....	.....	0.39	4.64	99.70	16.44	[−2.32−]		2.09	20.85	2

* Nomenclature according to A. L. Coulson, *Mem. Geol. Surv. India*, Vol. LXXV (1940), p. 1.

† Fe from FeO and Fe_2O_3.

‡ P from P and P_2O_5.

§ O from P_2O_5, FeO, and Fe_2O_3.

TABLE 1—*Continued*

Meteorite*	Silicate Phase																	Iron-Nickel Phase					Ref.
	SiO_2	MgO	Fe†	Al_2O_3	CaO	Na_2O	K_2O	Cr_2O_3	NiO	MnO	CoO	TiO_2	H_2O	P‡	O§	Total	Fe	Ni	Co	S	Total	
Khairpur, India	57.35	31.06	2.84	2.37	2.33	1.08	0.13	0.55	—	...	...	—	0.27	Tr.	1.20	99.18	21.36	1.29	0.17	2.84	25.66	6
Klein-Wenden, Germany	46.11	33.00	7.53	5.23	3.95	0.39	0.53	0.87	...	0.10	...	...	...	0.03	2.16	99.90	23.90	2.37	...	2.09	28.36	2
Knyahinya, Czechoslovakia	43.37	32.76	10.84	3.45	4.68	0.82	0.03	0.11	...	0.85	...	0.08	0.06	0.06	3.15	100.26	...	...	...	...	...	15
Krahenberg, Germany	47.86	21.67	15.76	3.75	2.40	0.20	1.42	1.04	...	0.91	...	...	...	0.54	4.52	100.07	10.37	1.36	...	2.35	14.08	2
Lanzenkirchen, Austria	46.64	27.59	11.48	4.98	1.94	2.68	0.32	0.51	...	0.08	...	0.08	0.06	0.10	3.32	99.78	11.48	1.10	0.06	1.90	14.54	16
Launton, England	46.41	29.10	12.44	2.81	2.44	1.18	0.11	0.36	0.14	0.34	...	0.21	0.49	0.14	3.78	99.95	11.00	1.20	0.08	2.18	14.46	6
Limerick Co., Ireland	48.53	30.69	10.17	1.97	2.45	1.25	0.18	0.78	0.25	0.20	...	...	0.89	0.27	3.00	100.63	20.38	1.51	0.12	2.03	24.04	7
Linum, Germany	54.93	32.82	1.31	3.11	4.45	1.77	0.33	0.40	...	0.26	...	...	0.15	0.10	0.37	100.00	19.06	0.71	...	1.85	21.62	2
Lissa, Czechoslovakia	43.12	29.41	14.57	2.25	3.02	1.66	0.46	0.07	...	1.12	...	0.02	0.04	0.17	4.35	100.26	...	...	...	...	...	15
Lixna, Latvia	46.00	31.64	13.11	3.18	Tr.	0.91	Tr.	0.56	...	1.03	...	...	...	0.18	3.75	100.36	16.95	1.71	...	2.12	20.78	2
Long Island, Kan.	37.61	23.99	18.66	3.25	1.48	0.26	0.03	6.68	0.81	Tr.	0.06	...	1.60	0.06	5.35	99.84	2.60	0.67	0.04	1.90	5.21	2
Lundsgard, Sweden	45.51	29.29	12.61	3.32	1.72	1.75	0.53	0.73	0.06	...	...	...	0.62	0.12	3.61	99.87	14.46	1.91	0.02	2.38	18.77	2
Lyubimovka, Russia	49.45	17.40	12.44	9.66	6.39	...	...	0.04	...	0.72	...	...	...	0.01	3.87	100.28	...	...	...	0.18	0.18	17
Macibini, S. Africa	49.98	8.58	13.04	11.92	10.50	0.63	0.06	0.44	...	0.52	...	0.75	0.13	0.07	3.75	100.37	1.58	Tr.	—	0.21	1.79	18
Manbhoom, India	43.09	29.32	17.76	1.93	2.07	0.47	0.21	0.59	...	0.08	...	...	...	0.21	5.09	100.82	4.24	0.91	...	1.70	6.85	2
Mangwendi, S. Rhodesia	44.00	26.03	16.88	4.72	1.68	0.50	0.12	0.49	0.37	0.15	...	0.21	0.12	0.17	4.88	100.32	4.91	0.92	0.11	1.83	7.77	19
McKinney, Tex.	43.21	30.43	6.55	15.15	1.88	...	...	0.08	0.50	0.24	...	...	...	0.06	1.88	99.99	9.05	0.92	0.05	2.28	12.30	2
Merva, India	49.68	31.80	7.80	3.41	2.45	0.79	0.09	0.50	...	0.24	...	0.18	0.47	0.29	2.57	100.07	21.78	1.82	0.11	2.07	25.78	20
Misshof, Russia	45.06	24.92	11.87	0.37	...	5.08	1.46	1.38	...	5.63	...	...	...	...	3.40	99.17	18.80	1.35	...	2.29	22.44	2
Mocs, Rumania	48.55	18.12	18.86	Tr.	3.16	1.36	0.24	1.20	...	1.92	...	...	...	0.47	5.40	99.28	7.93	1.38	Tr.	2.61	11.92	2
Modoc, Kan.	48.32	28.96	13.08	2.70	1.91	0.48	Tr.	...	...	0.11	...	...	...	0.05	3.75	99.36	6.56	0.68	0.03	1.38	8.65	2
Monroe, N.C.	44.42	29.69	13.92	4.34	2.75	—	—	Tr.	...	0.28	...	...	...	...	3.99	99.95	14.97	0.87	0.09	1.41	7.34	2
Moorleah, Tasmania	46.89	30.06	12.94	3.26	1.73	1.53	0.05	...	...	Tr.	...	...	0.06	Tr.	3.73	100.28	10.66	0.89	0.03	2.29	13.87	21
Mount Browne, Australia	43.72	29.33	12.21	2.85	2.81	1.47	0.30	0.03	...	Tr.	...	Tr.	...	0.14	3.50	96.36	16.33	1.78	0.26	2.02	20.39	2
Nakhla, Egypt	49.01	12.02	16.18	1.74	15.19	0.41	0.14	0.33	...	0.09	...	0.38	0.07	...	4.63	100.19	—	—	...	0.06	0.06	2
New Concord, Ohio	46.74	24.24	21.52	0.31	0.02	[1.03]		...	...	Tr.	...	...	...	...	6.17	100.03	9.23	1.31	0.04	0.11	10.69	2
Ngawi, Java	47.13	16.87	20.72	0.86	2.90	3.01	0.50	0.39	1.73	Tr.	...	...	...	...	5.93	100.04	6.50	0.65	Tr.	2.08	9.23	2
Novo-Urei, Russia	41.84	37.91	10.99	0.64	1.48	...	...	1.01	...	0.46	...	...	...	0.02	3.15	97.50	5.25	0.20	...	0.15	5.60	2
Oesede, Germany	47.95	31.33	8.37	3.65	3.99	1.78	0.59	...	...	...	...	...	...	...	2.40	100.01	21.83	1.90	...	2.11	25.84	22
Ojuelos Altos, Spain	47.70	30.25	10.51	3.30	2.97	0.91	0.27	...	...	...	...	...	0.11	0.31	3.11	99.44	12.46	1.12	0.04	2.47	16.09	23
Olivenza, Spain	43.91	27.12	17.38	2.48	2.25	0.99	0.36	0.15	...	0.16	...	—	...	0.15	5.03	99.98	4.59	0.41	0.02	2.19	7.21	24
Olmedilla de Alarcon, Spain	48.14	30.76	9.51	3.29	2.82	1.13	0.25	0.50	...	0.34	...	—	0.12	0.30	2.83	99.99	20.02	1.58	0.02	1.70	23.32	25
Ferpeti, India	45.73	29.38	13.62	1.58	2.29	1.37	0.16	0.76	...	0.32	...	0.19	...	0.24	4.48	100.12	10.89	1.18	0.05	2.28	14.40	26
Phu-Long, Cochin China	46.26	29.62	11.78	4.18	2.82	0.91	0.27	0.14	...	0.17	...	...	0.08	0.20	3.44	99.87	18.68	1.21	0.02	1.65	21.56	27
Plantersville, Tex.	48.09	29.63	11.82	2.74	2.48	1.32	0.15	0.04	0.01	0.23	—	—	...	0.05	3.40	96.96	16.99	1.92	0.06	1.45	20.42	28
Prambachkirchen, Austria	46.37	28.25	11.66	4.67	2.70	1.40	0.28	0.06	...	1.46	...	0.14	...	0.29	3.43	100.71	13.20	0.88	0.07	2.80	16.95	29
Rangala, India	46.82	28.86	13.67	2.66	2.42	0.71	0.16	0.34	...	0.14	...	0.09	0.04	0.23	4.04	100.18	11.52	1.00	0.14	1.86	14.52	30
Renca, Argentina	43.34	22.64	18.24	3.42	1.63	2.68	0.65	0.70	1.28	0.88	—	—	—	0.02	5.24	100.72	7.79	0.79	0.05	2.80	11.43	31
Rochester, Ind.	46.55	30.33	14.74	0.11	2.77	0.65	...	0.11	...	...	...	...	...	...	4.22	99.48	11.40	0.41	...	1.09	12.90	2
Roda, Spain	51.29	27.59	11.80	2.91	1.45	0.39	0.32	0.65	...	0.22	...	—	0.51	0.03	3.39	100.55	1.10	...	...	0.63	1.73	32
Rosebud, Tex.	41.17	28.55	17.54	4.24	2.16	0.49	0.07	0.17	0.17	0.18	Tr.	—	0.08	0.08	5.06	99.96	14.32	1.15	0.05	1.25	16.72	33
St. Christophe, France	47.08	31.00	12.71	2.57	1.81	0.61	0.22	0.47	...	...	...	...	...	...	3.64	100.11	12.17	1.67	0.11	2.52	16.47	2
St. Marks, S. Africa	57.78	27.51	7.63	0.97	2.20	1.28	0.35	...	...	1.06	...	...	...	0.08	2.18	101.04	26.44	1.84	0.21	5.26	33.75	2
St. Michel, Finland	46.63	29.03	12.33	3.91	1.94	1.56	0.15	0.66	...	0.48	...	0.02	...	0.09	3.53	100.33	11.71	1.16	0.13	2.22	15.22	2
Saline, Kan.	41.46	26.09	19.15	2.05	2.27	0.29	0.09	1.40	0.83	...	0.08	...	1.38	0.06	5.49	100.64	7.89	0.95	0.04	1.65	10.53	2
Salt Lake, Utah	46.61	29.76	12.28	2.53	2.42	1.10	0.08	0.54	...	...	...	...	1.22	0.26	3.60	100.40	19.18	1.38	0.10	2.00	22.66	2
Santa Isabel, Argentina	47.22	23.32	15.79	3.37	2.92	0.76	0.27	0.76	1.15	0.45	Tr.	Tr.	...	0.16	4.59	100.76	8.37	0.30	0.02	2.83	11.61	34

TABLE 1—*Continued*

Meteorite*	Silicate Phase																Iron-Nickel Phase					Ref.
	SiO_2	MgO	Fe†	Al_2O_3	CaO	Na_2O	K_2O	Cr_2O_3	NiO	MnO	CoO	TiO_2	H_2O	P‡	O§	Total	Fe	Ni	Co	S	Total	
Sazovice, Czechoslovakia	45.87	27.73	13.84	3.68	2.83	1.03	.....	.....	.....	0.83	.....	0.07	0.10	0.05	3.97	100.00	.....	.....	.....	.....	.....	35
Schoneberg, Germany	48.44	16.67	16.24	6.72	2.79	2.66	0.88	0.49	.....	.....	.....	.....	.....	0.43	4.65	99.97	13.77	1.47	.....	1.93	17.17	2
Sharps, Va.	43.24	28.69	14.46	3.33	3.05	0.94	0.21	.....	0.48	0.36	—	.....	.....	0.17	4.17	99.15	.....	.....	.....	0.58	0.58	36
Shelburne, Canada	45.11	30.20	13.56	2.47	2.01	0.84	0.25	0.71	.....	0.14	.....	.....	.....	0.07	3.89	99.25	10.70	0.78	0.04	1.61	13.13	2
Soko-Banja, Yugoslavia	45.72	28.27	15.20	2.34	2.12	0.94	0.08	0.40	.....	0.06	.....	0.11	0.19	0.20	4.42	100.05	6.34	1.06	0.09	1.81	9.30	6
Stalldalen, Sweden	47.71	30.94	10.69	2.82	2.15	0.83	0.20	0.53	0.27	0.33	.....	.....	.....	0.19	3.06	99.72	21.10	1.61	0.17	2.27	25.15	2
Stannern, Czechoslovakia	48.37	7.20	14.88	11.29	10.45	0.75	0.13	0.35	0.25	.....	Tr.	0.41	0.30	0.11	4.29	98.79	0.55	.....	.....	0.31	0.86	2
Strathmore, Scotland	47.38	29.33	12.15	3.02	2.16	1.06	0.13	0.52	.....	0.34	.....	0.16	0.20	0.20	3.59	100.24	11.31	1.26	0.05	2.29	14.91	37
Tabor, Czechoslovakia	43.89	32.10	11.55	2.91	3.82	1.25	0.09	0.10	.....	0.71	.....	0.09	0.10	0.14	3.43	100.18	.....	.....	.....	.....	.....	15
Tataovine, Southern Tunis	55.16	27.53	11.94	0.62	0.76	.....	—	0.35	—	0.26	.....	0.19	0.14	.....	3.42	100.37	0.22	.....	.....	0.13	0.35	38
Tennasilm, Esthonia	45.29	25.91	15.86	2.97	1.59	[2.61]		1.07	.....	.....	.....	.....	.....	0.08	4.54	99.92	11.77	1.67	.....	0.63	14.07	2
Tilden, Ill.	44.72	24.20	18.79	4.00	0.99	0.58	—	.....	1.07	.....	0.04	.....	0.03	0.12	5.42	99.96	6.38	0.31	0.02	1.15	7.86	39
Tjerebon, Java	43.22	26.06	20.12	.....	0.97	0.91	0.52	0.74	0.16	0.57	.....	.....	.....	.....	6.05	99.32	14.21	1.49	.....	1.99	17.69	40
Travis County, Tex.	46.09	28.77	12.84	2.80	2.30	1.16	0.13	0.54	0.54	Tr.	.....	.....	0.87	0.32	3.78	100.14	1.80	0.22	0.01	0.91	2.94	2
Tuantuc, Cochin China	46.36	28.94	13.33	3.41	2.29	0.77	0.24	0.16	.....	0.24	.....	.....	0.07	0.25	3.90	99.96	9.93	0.78	0.02	2.24	12.97	27
Utrecht, Holland	45.93	28.42	14.05	2.61	1.73	1.62	0.17	0.51	.....	.....	.....	.....	.....	Tr.	4.02	99.06	11.07	1.24	.....	1.94	14.25	2
Wittekrantz, S. Africa	46.67	28.83	13.18	2.88	2.41	1.32	0.16	0.41	.....	0.17	.....	0.19	0.18	0.14	3.82	100.36	9.85	0.75	0.02	1.26	11.88	2
Zebrak, Czechoslovakia	43.04	32.14	12.45	2.85	3.05	1.54	0.33	0.08	.....	1.07	.....	0.05	0.06	0.09	3.65	100.40	.....	.....	.....	.....	.....	15
Zemaitkiemis, Lithuania	48.80	27.04	11.33	3.73	2.78	2.49	0.31	.....	.....	0.27	.....	0.06	.....	0.27	3.36	100.44	14.74	0.80	.....	1.90	17.44	41

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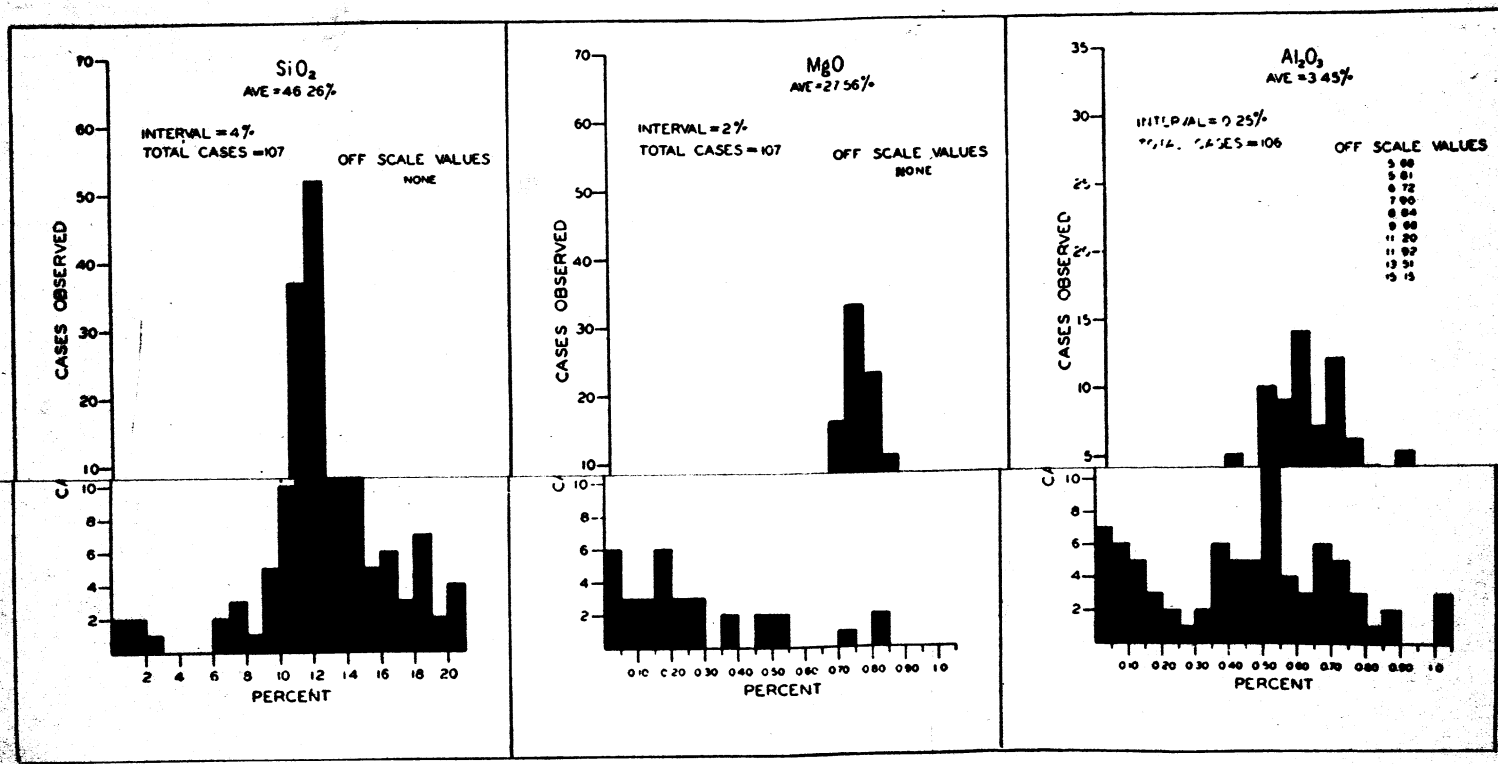

FIG. 1.—Frequency-distribution curves for the major constituents of the silicate phase of stony meteorites

Merrill's compilation, and the balance have been selected from the literature accumulated since 1916. The data have been broken down in the following manner: The metallic iron, nickel, and cobalt and the sulphur present in each meteorite have been subtracted from the gross analysis. The balance, representing the silicate phase, has been adjusted so that the sum of the constituents equals 100 per cent. Wherever possible the direct analysis of the silicate phase, freed magnetically from the metallic and sulphide phases, was used, in which case no numerical adjustments were necessary.

Figure 1 shows frequency-distribution curves for the major constituents. The number of observed cases lying within a designated percentage interval is plotted against the location of the interval. The interval chosen in each case is approximately 10 per cent of the arithmetic mean. It can be seen that the two most abundant constituents, silica and magnesia, are fairly well defined and that for the less abundant constituents the spread is considerably greater.

Table 2 shows the arithmetic mean of each constituent, together with the standard deviation σ, where

$$\sigma = \sqrt{\frac{\Sigma\Delta_i^2}{n-1}};$$

Δ_i = Deviation of the abundance of a given constituent in a given meteorite from the arithmetic mean;

n = Number of meteorites studied.

The standard deviation is essentially a measure of the "spread" of an element or the range of concentrations in which it is most prevalent. If a sufficient number of cases is available for the data to be representative, σ should be fairly independent of n.

TABLE 2

AVERAGE COMPOSITION BY WEIGHT OF THE SILICATE PHASE OF STONY METEORITES

Constituent	Per Cent	Standard Deviation	Precision	Observed Cases
SiO_2	46.26	4.18	± 0.40	107
MgO	27.56	5.44	± 0.53	107
Fe*	13.25	4.39	± 0.42	107
Al_2O_3	3.45	2.92	± 0.28	106
CaO	2.90	2.96	± 0.29	106
Na_2O	1.10_3	0.75	$\pm 0.07_5$	99
K_2O	0.25_3	0.25	$\pm 0.02_6$	94
Cr_2O_3	0.54_2	0.73	$\pm 0.07_8$	89
	0.47_2	0.33	$\pm 0.03_5$	88
MnO	0.41_5	0.71	$\pm 0.07_8$	82
	0.35_1	0.36	$\pm 0.03_9$	81
NiO	0.50_2	0.58	$\pm 0.09_7$	41
TiO_2	0.17_3	0.37	$\pm 0.05_3$	48
	0.12_5	0.16	$\pm 0.02_3$	47
CoO	0.027	0.046	± 0.011	16
H_2O	0.58_5	1.00	$\pm 0.14_3$	49
P	0.17	0.13	± 0.14	85
O†	3.84	1.29	± 0.12	107

* From Fe_2O_3 and FeO.
† From Fe_2O_3, FeO, and P_2O_5.

The precisions of the arithmetic means are given in the fourth column of Table 2 and have been calculated, using the relationship

$$M = \pm\frac{\sigma}{\sqrt{n}},$$

where M is the measure of precision. In the cases of the three constituents—titanium oxide, chromium oxide, and manganese oxide—the arithmetic means are affected markedly by the decision as to whether to include in each case one abnormally high value. For these cases the means, together with the deviations and precisions, were calculated both with and without the abnormal value.[4]

In Table 3 the composition has been broken down in terms of elements and expressed in three ways: weight per cent, atomic per cent, and (after Goldschmidt) in terms of atoms per 100 atoms of silicon. In the cases of titanium, chromium, and manganese the averages given are the means of the high and low values.

[4] The abnormal values are as follows: Angra dos Reis, Brazil, 2.42 per cent TiO_2; Long Island, Kansas, 6.68 per cent Cr_2O_3; Misshof, Russia, 5.63 per cent MnO.

It can be seen that in the most favorable case, that of silicon, the mean is probably correct to within 1 per cent of the mean. In the least favorable case, that of titanium, the mean is probably correct to within 35 per cent of the mean.

If one plots the standard deviation divided by the arithmetic mean against the logarithm of the arithmetic mean for each of the metallic oxides, an interesting trend becomes evident. Figure 2 indicates that, as the concentration decreases, the standard deviation increases rather markedly. While at least some of the observed spread can be attributed to experimental error, it seems clear that the major portion of the spreads, shown in the plot, actually occur in nature. While the curve, arbitrarily drawn through the points, should not be regarded as a precise relationship, Figure 2 indicates that in the case of the minor constituents (constituents present to an extent less than 100 parts per million) a very large number of falls must be analyzed if arithmetic means possessing fair degrees of precision are to be

TABLE 3

AVERAGE COMPOSITION OF THE SILICATE PHASE OF STONY METEORITES BY ELEMENTS

Element	Weight Per Cent	Atomic Per Cent	Atoms per 100 Atoms of Si
Oxygen	43.10 ±1.00	58.09 ±1.36	350 ±11
Silicon	21.61 ±0.19	16.60 ±0.15	100
Magnesium	16.62 ±0.32	14.73 ±0.28	88.7 ± 2.5
Iron	13.25 ±0.42	5.11 ±0.16	30.8 ± 1.3
Calcium	2.07 ±0.21	1.11 ±0.11	6.70 ± 0.74
Aluminum	1.83 ±0.15	1.46 ±0.12	8.82 ± 0.81
Sodium	0.81_8 ±0.05_6	0.76_8 ±0.05_2	4.62 ± 0.36
Nickel	0.39_4 ±0.07_2	0.14_5 ±0.02_5	0.87 ± 0.16
Chromium	0.36_2 ±0.06_2	0.15_0 ±0.02_5	0.90 ± 0.16
Manganese	0.31_1 ±0.07_0	0.12_2 ±0.02_5	0.74 ± 0.16
Potassium	0.20_9 ±0.02_1	0.11_5 ±0.01_1	0.69_3 ± 0.07_5
Titanium	0.098±0.032	0.04_3 ±0.01_4	0.26_0 ± 0.09_0
Phosphorus	0.16_6 ±0.01_4	0.11_6 ±0.01_0	0.69_6 ± 0.06_5
Hydrogen	0.066±0.016	1.40 ±0.35	8.4 ± 2.2
Cobalt	0.022±0.009	0.0079±0.0033	0.048± 0.020

obtained. For example, extrapolation of the curve in Figure 2 back to 10 parts per million (0.001 per cent) indicates that, for a constituent present at that average concentration, some 500 falls should be analyzed (either collectively or separately) if the arithmetic mean is to be precise to 10 per cent.

no greater a precision than 25–35 per cent on a purely statistical basis. Such a precision, however, would be correct only if no error is inherent in the analytical work. When one takes into consideration the difficulties of precise analysis, it seems only reasonable to suppose that most of the work done to date on the

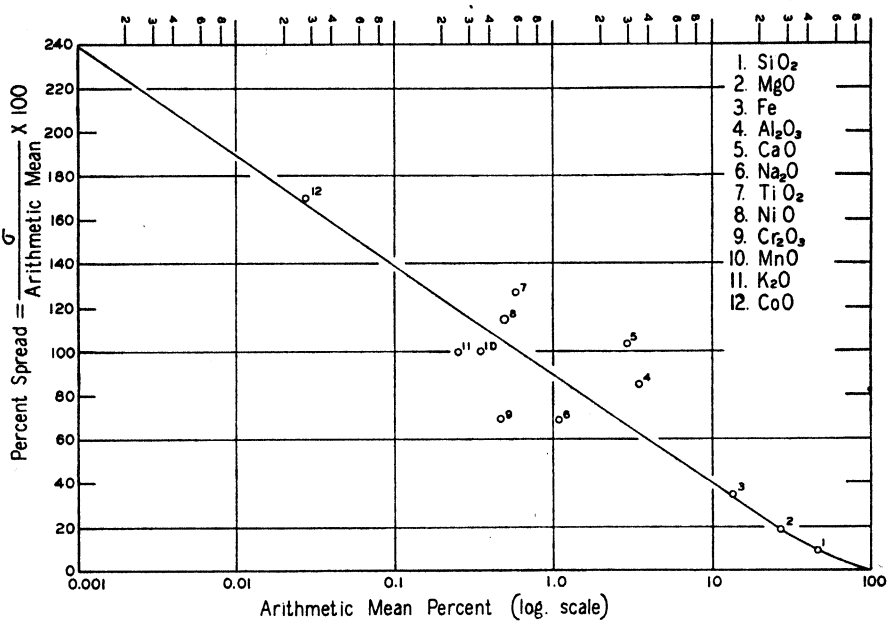

FIG. 2.—The variation of the spread with the arithmetic mean for the major constituents of the silicate phase of stony meteorites.

Thus it would appear that in the extensive work of the Noddacks[5] on the minor constituents of stony meteorites (a composite mixture of 42 falls was used), the results are probably valid to

minor constituents of stony meteorites has a precision of probably no greater than a factor of 2.

ACKNOWLEDGMENTS.—The authors are indebted to Professors T. Barth and Jerome Fisher, of the Department of Geology of the University of Chicago, and to Dr. S. K. Roy, of the Chicago Natural History Museum, for many stimulating conversations on the general subject of this paper.

[5] I. and W. Noddack, "Die Häufigkeit der chemischen Elemente," *Naturw.*, Vol. XVIII (1930), p. 757; "Die geochemischen Verteilungskoeffizienten der Elemente," *Svensk kem. Tidssk.*, Vol. XLVI (1934), p. 173.

THE LONGVIEW MEMBER OF THE KINGSPORT FORMATION

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Emory University

ABSTRACT

A detailed stratigraphic study of the Longview member of the Kingsport formation of the Knox dolomite group has been undertaken because of the economic importance of the Knox dolomite to zinc mining in the southeastern United States.

An excellent opportunity presented itself to the author to study and correlate detailed sections in underground exposures of zones about which little could be learned from surface exposures because of the scarcity of outcrops, the variation in lithology, and the lack of fossils.

INTRODUCTION

The Mascot properties of the American Zinc, Lead, and Smelting Company are in Knox County, Tennessee, 13 miles northeast of Knoxville, about equidistant between U.S. Highways 11*E* and 11*W*. The data for this paper were gathered from a study of underground exposures in the Mascot mines and from surface outcrops on the company property and near-by localities.

The mineralized district lies within the Tennessee section of the Valley and Ridge Province, a southwest extension of the Shenandoah Valley of Virginia. This part of the Valley and Ridge Province is approximately 50 miles wide, being bounded on the northwest by the Cumberland Plateau and on the southeast by the Great Smoky Mountains. Topographically, it is a region of ridges with narrow intervening valleys. The ridges are formed of resistant chert and sandstone, and the valleys of less resistant limestone, dolomite, and shale, The ridges are largely wooded. The valleys are used for agriculture and pasture land.

HISTORY OF ZINC MINING IN TENNESSEE

Zinc mining in eastern Tennessee dates back to the middle of the last century. Oxide and sulphide ores occur at many places. Where conditions were favorable for their formation, carbonate ores occurred in the outcrop, from which the earliest production was derived.

Tennessee first came to general attention as an important producer of zinc ores in 1913, when the Embree Iron Company shipped its first oxidized ores from Embreeville and the American Zinc Company its first sulphide ores from Mascot.

The Mascot–Jefferson City zinc district extends from Mascot to Jefferson City, about 30 miles northeast of Knoxville. The principal producers of sulphide ores at the present time are the American Zinc Company mines at Jefferson City, New Market, and Mascot; and the Universal Exploration Company mine at Jefferson City. The approximate daily production is 5,000 tons of ore, grading 2.5–5 per cent zinc.

Zinc ores occur at various horizons in the Knox dolomite. The position of the most productive ores at Mascot has been a zone 97 feet thick, occurring some 150 feet above the top of the Longview member; whereas the ore mined at New Prospect, 25 miles northwest of Mascot, is 2,500 feet stratigraphically lower. Since the Longview occupies an intermediate position and in some localities is known to contain zinc minerals, a detailed description of it should be useful.

GENERAL GEOLOGY OF THE DISTRICT

The formations of the district are of limestones, dolomites, and shales, ranging in age from Cambrian to Carboniferous.[1] The strata have been intensely folded and faulted, the structures being essentially parallel and trending northeasterly with the valley. The prevailing rock dips in this region of Appalachian overturning are to the southeast. Displacements along some of the overthrusts amount to as much as 6–8 miles. The folding and faulting have allowed the unequally resistant strata to form alternate ridges and valleys and to outcrop in a succession of belts.

[1] M. H. Newman, "The Mascot–Jefferson City Zinc District, *XVI Int. Geol. Cong. Guidebook 2* (1932), pp. 152–65.

The Knox dolomite, which ranges in thickness from 2,800 to 3,200 feet, is the most important and the most widespread of the valley rocks. It is made up of at least four formations, which are so similar in appearance that very close study is required to distinguish them. These are the Conococheague, Chepultepec, Kingsport, and Mascot. Their stratigraphic position, subdivisions, and approximate thicknesses are shown in the accompanying generalized section (Table 1).

The Kingsport formation of the Knox group is composed of 550–625 feet of limestone and dolomite. The dolomite varies in color from very light to dark gray and in texture from fine grained to very coarse grained. The limestone is

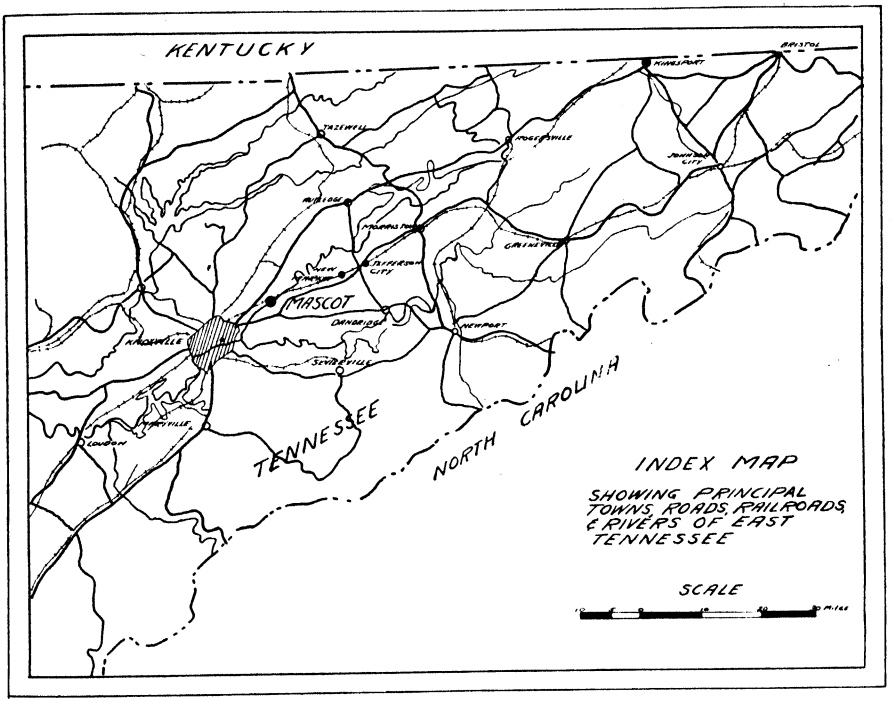


FIG. 1

dove to brown in color and is fine grained. Numerous arenaceous layers and shale seams are present, but few are of very great thickness. Many bands of chert nodules and chert layers are scattered throughout the section. The lowest divison of the Kingsport is the Longview dolomite member.

STRATIGRAPHY OF THE LONGVIEW DOLOMITE

In older publications the portion of the Knox dolomite represented by the Longview dolomite member was called "Nittany" because it was believed to correlate with a formation bearing that name in Pennsylvania. Recent workers, how-

TABLE 1

GENERALIZED SECTION, MASCOT–JEFFERSON CITY ZINC DISTRICT, EASTERN TENNESSEE

Age		Formation Name			Thickness in Feet	Character of Rocks
Ordovician	Chazyan	Lenoir and Mosheim limestone (Chickamauga)			250±	Dark gray (Lenoir) and dove bird's-eye (Mosheim) limestone
Ordovician	Canadian	Knox dolomite	Mascot (Cotter-Powell)		500–700	Light and dark-gray dolomite and limestone moderately cherty; limestone most abundant in Jefferson City area; base marked by chert-matrix sandstone; rarely mineralized
Ordovician	Canadian	Knox dolomite	Kingsport (Jefferson City)	Upper dolomite	184±	Very light- to medium light-gray dolomite, moderately cherty; extremely fine-grained with small amount semicrystalline and crystalline dolomite; base marked by 5-foot bed of very dark dolomite
Ordovician	Canadian	Knox dolomite	Kingsport (Jefferson City)	Middle limestone	210±	Brown limestone, which in many places has been altered to crystalline dolomite; some some light- to dark-gray dolomite
Ordovician	Canadian	Knox dolomite	Kingsport (Jefferson City)	Longview dolomite	150±	Dolomite, light gray to nearly white; fine-grained to semicrystalline dolomite
Cambrian	Croixian	Knox dolomite	Chepultepec		500±	Mostly dolomite at Mascot; light- and dark-gray dolomite and limestone in Jefferson City area; very cherty; heavy sandstone at base
Cambrian	Croixian	Knox dolomite	Conococheague		500±	Dark sugary and light-gray, fine-grained cherty dolomite; some limestone in Jefferson City area
Cambrian	Croixian	Maynardsville limestone			60–250+	Light-blue to dark-gray laminated limestone and dolomite
Cambrian	Croixian	Nolichucky shale			550–750	Dark-gray, greenish, and yellow shales and thin limestones

ever, have replaced the name Nittany with that of "Longview," which was first used by Charles Butts[2] in describing rocks in Alabama which are now believed to correlate with this zone in Tennessee.

The Knox dolomite has been described by various authors, but, with the exception of the descriptions of the Kingsport formation by C. R. L. Oder and W. H. Miller[3] and by J. Crawford,[4] practically all the descriptions have dealt with surface outcrops involving, for the most part, poor exposures of the Longview.

The detailed stratigraphys and the true relations of the Longview have heretofore been poorly understood because surface exposures of the rocks in the district are few. Furthermore, fossils are sparse and difficult to find, and the ones that can be found occur in residual chert. Lithology appears to be more practicable than fossils for establishing the the Longview as a mappable unit and for understanding its nature and relation to the Kingsport formation. Since the parts of this unit seen in underground workings at Mascot are more completely exposed than in any other locality in the

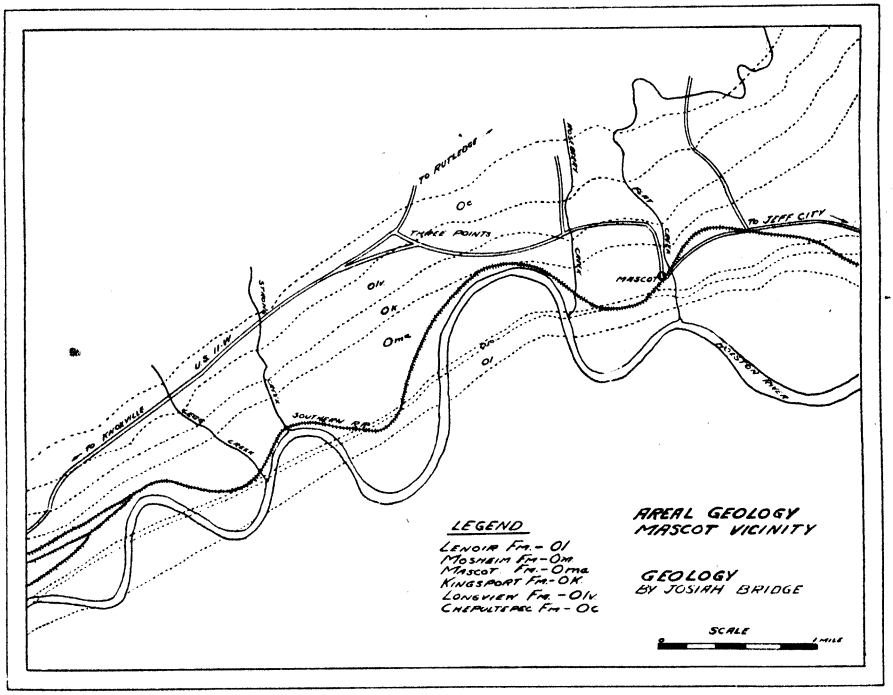


FIG. 2

[2] "Geology of Alabama, the Paleozoic Rocks," *Ala. Geol. Surv. Spec. Rept. 14* (1926).

[3] "Stratigraphy of the Mascot–Jefferson City Zinc District," *A.I.M.E. Tech. Pub. 1818* (1945), pp. 1–9.

[4] "Structural and Stratigraphic Control of Zinc Deposits in East Tennessee," *Econ. Geol.*, Vol. XL, No. 6 (1945), pp. 412–15.

valley, except in the section at Thorn Hill, the mine sections can best furnish the detailed descriptions.

The Longview member of the Mascot district is composed of 145 feet of light- to very-light-gray dolomite, for the most part in fine-grained layers, although some layers are so coarsely crystalline that they resemble marble in appearance. The rock approaches theoretical dolomite in composition and may have been altered from fine-grained limestone. Some layers show a replacement of the limestone by crystalline dolomite within only a few feet in underground exposures. Arenaceous zones of medium-sized frosted quartz grains disseminated in dolomite are numerous, although most of the silica content occurs as interstitial filling. Chert nodules and chert layers are scarce.

The late Mark H. Newman devised the method of measuring and naming the beds in the mines at Mascot and described parts of the ore-bearing zones. His method utilized numbers to denote the thickness of beds above and below the footwall sand, which he designated the "0" bed. These numbers also were used as the name of the individual bed.

Oder and Miller's studies did not include any of the Longview. In fact, they did not cover any of the beds between the top of the Longview and the layer which they termed the "−64 bed" at Mascot. Newman recognized the presence of sands below −64 bed, but he failed to describe any part of that zone.

SUMMARY

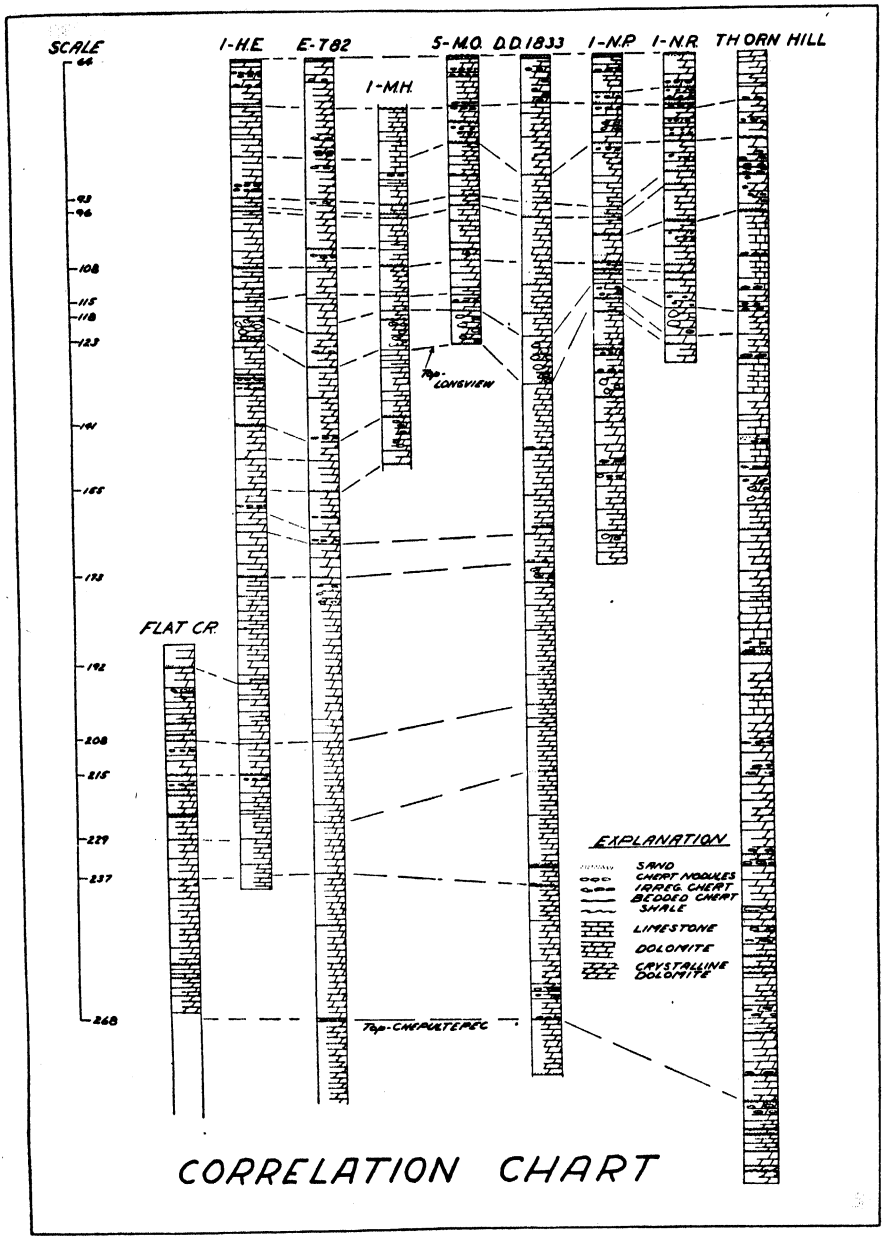
Former workers have considered the thickness of the Longview in the Mascot locality to be 250 feet. The writer's examination of the strata included in that measurement shows that it includes about 100 feet of crystalline dolomite and layers of true limestone, which belong in the overlying middle limestone member of the Kingsport.

The writer has found no radical change in sedimentation or other evidence of unconformity at the top of the Longview in this area. The lithology of that unit is more like that of the overlying Kingsport than that of the underlying Chepultepec beds. Therefore, it is here considered to be a member of the Kingsport formation, which the author believes consists of three members as follows:

1. The upper dolomite member; 184 feet thick; principally light-gray, fine-grained dolomite, with a few thin layers of limestone
2. The middle limestone member; 210 feet thick; composed of brown limestone, which in many places has been altered to crystalline dolomite with some layers of light- to dark-gray, fine-grained dolomite
3. The Longview dolomite member; 145 feet thick; composed of light-gray to nearly white, fine-grained dolomite, with a few beds of light-gray dolomite of medium crystalline texture.

The crystalline dolomite in the lower part of the middle member of the Kingsport formation is referred to locally as "recrystalline dolomite." When one follows this zone along the strike of mine sections, the crystalline texture may be seen to grade into unaltered, fine-grained, dove-colored limestone. In some localities dolomitization and the development of a crystalline texture are practically complete, with only a few remnants of unaltered limestone remaining. The evidence noted in this zone is sufficiently conclusive to state that the crystalline texture is a result of alteration.

Whether or not any limestone occurs in the Longview has been a matter of controversy. It is certainly present in this zone in the Thorn Hill section, but only one small occurence of actual lime-


SCALE
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FLAT CR.
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E-T82
1-M.H.
5-M.O.
D.D.1833
1-N.P
1-N.R
THORN HILL
Top-LONGVIEW
Top-CHAPULTEPEC
EXPLANATION
SAND
CHERT NODULES
IRREG. CHERT
BEDDED CHERT
SHALE
LIMESTONE
DOLOMITE
CRYSTALLINE DOLOMITE
CORRELATION CHART

FIG. 3

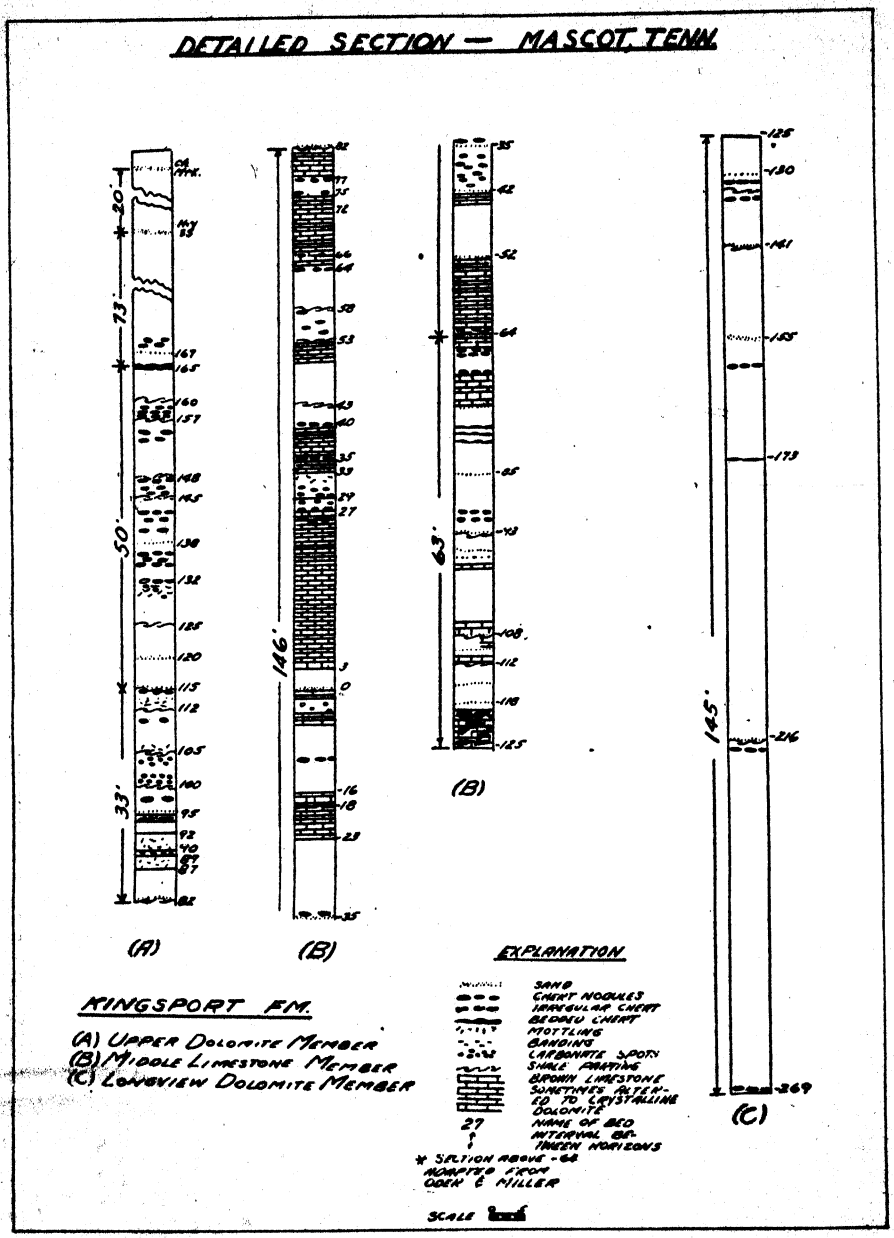
DETAILED SECTION — MASCOT, TENN.
KINGSPORT FM.
(A) UPPER DOLOMITE MEMBER
(B) MIDDLE LIMESTONE MEMBER
(C) LONGVIEW DOLOMITE MEMBER
EXPLANATION
SAND
CHERT NODULES
IRREGULAR CHERT
BEDDED CHERT
MOTTLING
BANDING
CARBONATE SPOTS
SHALE PARTING
BROWN LIMESTONE
SOMETIMES ALTERED TO CRYSTALLINE DOLOMITE
NAME OF BED
INTERVAL BETWEEN HORIZONS
* SECTION ABOVE -68 MODIFIED FROM ODEN & MILLER
SCALE

FIG. 4

stone has been found in the Longview at Mascot, although some of the crystalline dolomite occurring in this member in the latter locality represents altered limestone.

Some workers have thought that chert is far more abundant in the Longview than in other parts of the Kingsport formation. This opinion seems to be due to the heavy chert residuum usually found along the surface outcrop of this zone. However, the writer has found that this is not always true. If the limestone and crystalline dolomite members are placed in the Longview, because of the presence of *Lacanospira*, the Longview may be considered to be as cherty as the middle member; but if the limestone beds are placed in the middle member of the Kingsport, the amount of chert in the Longview, other than interstitial silica, is small in quantity.

ACKNOWLEDGMENTS.—The writer wishes to express his appreciation to the American Zinc Company, especially to Mr. Harley A. Coy, general superintendent, for permission to do this work and to Dr. Charles R. L. Oder, chief geologist, for advice and many helpful suggestions.

ON THE GEOCHEMICAL CYCLE OF FLUORINE

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University of Chicago

ABSTRACT

The average concentration of fluorine in the earth's crust is about 800 gm. per ton. In basalts and gabbros it is present as apatite; in granites, gneisses, and metamorphic rocks it is contained in mica, hornblende, and fluorspar. Much fluorine enters into the magmatic gas phase, which has an acid reaction. Great amounts of fluorine are therefore exhaled by volcanic eruptions. If the magmatic gas phase has opportunity to react with the wall rock, it turns alkalic, and fluorine is extracted (as phosphate?) before it reaches the surface; alkaline hot springs carry no magmatic fluorine. Of the fluorine dissolved by the weathering processes, only 0.2 per cent is found in the ocean. Fluorine is extracted from inland waters and from the sea, forming mineral deposits with a composition resembling fluorapatite.

THE AVERAGE FLUORINE CONCENTRATION IN THE LITHOSPHERE

Geochemistry is concerned with the chemical composition of the earth's crust as a whole, with the differences in composition between the individual rocks and minerals, and with chemical changes occurring in the rocks. It traces the geochemical cycle of each element and explains reactions taking place at the various depths, e.g., how some elements in certain stages of their geochemical cycle concentrate in ore deposits or salt layers, while others during their entire cycle occur as "disperse" elements in very small concentrations.

Fluorine is one of these "disperse" elements, never concentrating in large mineral deposits, and, in addition, it has been difficult to catch analytically in rocks and minerals.

In 1920, F. W. Clarke[1] on very scanty data estimated fluorine to constitute about 200 gm. per ton of the earth's crust. Four years later F. W. Clarke and H. S. Washington[2] were so discouraged by the uncertainty of the analytical methods available for determining fluorine that they virtually neglected the data of the analysts and estimated the amount of fluorine indirectly, on the assumption that the chief carrier of fluorine in igneous rocks was the mineral apatite. Consequently, they argued, the average fluorine should be about one-tenth that of the phosphoric acid, i.e., 300 gm. per ton. This figure is now currently used in geochemical writings.

But with the publication of H. H. Willard and O. B. Winter's[3] reliable method for determining fluorine, E. S. Shepherd[4] at the Geophysical Laboratory started a long series of analyses aiming to establish the significance of fluorine in geochemistry.[5] He found that about 400 gm. per ton is characteristic of plutonic rocks, although alkaline rocks seemed to average higher. His data are interesting also in showing that fluorine does not, as had been suggested, increase in proportion to phosphorus—thus apa-

[1] "The Data of Geochemistry," *U.S. Geol. Surv., Bull. 695* (1920).

[2] "The Composition of the Earth's Crust," *U.S. Geol. Surv. Prof. Paper 127* (1924).

[3] "Volumetric Method for Determination of Fluorine," *Indust. Eng. Chem., Anal. Ed.*, Vol. V (1933), pp. 7–10.

[4] "Volatile Constituents of Rocks: Fluorine," *Carnegie Inst. Wash. Year Book No. 34* (1934–35), pp. 98–100.

[5] "The Gases in Rocks and Some Related Problems," *Amer. Jour. Sci.*, Vol. XXXVA (1938), pp. 311–51; "Note on the Fluorine Content of Rocks and Ocean Bottom Samples," *ibid.*, Vol. CCXXXVIII (1940), pp. 117–28.

tite is not the chief carrier of fluorine in igneous rocks.

At the annual meeting of the American Mineralogical Society, in December, 1946, Dr. B. Wasserstein read a paper entitled "On the Geochemistry of Fluorine." From a great many data he concluded that the fluorine concentration in granite (presumably including the extensive pre-Cambrian granites and gneisses) would be 800 gm. per ton or more. This figure is twice as high as that given by Shepherd for plutonic rocks; and, since the continental granite areas exceed those of all other plutonic rocks by a factor of 20, it would seem very probable that the average concentration of fluorine in the near-surface rocks is close to 800 gm. per ton. This means that fluorine would be number 10 in amount among all elements, following titanium in the list. Even in weight percentage it would rank higher than chlorine, phosphorous, and sulphur; in terms of atom percentages it would be three times as abundant as chlorine and between two and three times as abundant as manganese.

If we accept this estimate of the fluorine content of the earth's crust, the next questions are to find out how fluorine behaves; in which reactions it takes place; in which minerals it is, or can be, present; to what extent differences of concentrations occur and why they occur. In order to answer these questions, we must find the relation between fluorine in the lithosphere and in the sea.

FLUORINE IN SEA WATER

The salt in the sea, its origin, and the manner of its delivery into the sea, have not been satisfactorily explained. Most widely accepted is the theory that the sea was originally fresh and gradually became saline through addition of salts contained in river water.

C. H. White,[6] on the other hand, believes that the primeval sea was saline; that volatile alkaline chlorides were formed prior to and above the silicates; and that they came to rest upon the primeval silicate crust and remained there until they were dissolved by the early rains and were carried as saturated brine into the growing oceans.

J. Joly[7] has held the view that the primeval atmosphere contained HCl which, mixed with rain water, was able to dissolve metals from the silicate crust and carry them as chlorides into the sea, which originally was a dilute solution of HCl without sodium. Likewise, Joly has shown that, for each kilogram of sea water, about 600 gm. of average igneous rocks would have to pass through the cycle of weathering and sedimentation in order to deliver the necessary sodium, assuming that one-third of the soda of the parent-rock had been locked up in sediments, while two-thirds still remained in the sea. Although recognizing the uncertainties involved, we shall use Joly's figure as the basis for calculating the amounts of mineral matter that have been supplied to the sea. Even if this estimate is incorrect, there should probably be no great change in the proportional relations of the figures presented in Table 1.

Table 1 demonstrates the very interesting fact that chlorine and bromine are found in sea water in such large quantities that—if the data underlying the computations are anywhere near right—a derivation from the lithosphere

[6] "Why the Sea Is Salt," *Amer. Jour. Sci.*, Vol. CCIV (1942), 714–24.

[7] "An Estimate of the Geological Age of the Earth," *Sci. Trans. Roy. Soc. Dublin*, Vol. VII (1899), pp. 23–66.

through weathering processes seems impossible. The amount of chlorine thus delivered to the sea through all geologic ages would be 290 mg. per kilogram of sea water. But 1 kg. of sea water today contains 19,300 mg., or 6,655 per cent of the chlorine thus delivered. The percentages given in the last column of Table 1 should, for each element, represent the fraction of the total amount delivered that was retained in the sea. Consequently, figures higher than 100 per cent indicate that the corresponding elements must have been present as primary constituents of the sea or else were delivered to the sea by volcanic gases and hot springs.

TABLE 1

HALOGENS PRESENT IN ROCKS AND IN SEA WATER*

	Present in Rocks (Gm/Ton)	Delivered to Each Kg. of Sea Water by Weathering Processes (Mg.)	Present in Sea (Mg. per Kg. of Water)	Percentage
F.....	800	480	1	0.2
Cl.....	480	290	19300	6655
Br.....	2	1.2	66	5500
I.....	0.3	0.18	0.05	28

* The values for Cl, Br, and I in rocks are taken from: F. W. Clarke and H. S. Washington, "The Composition of the Earth's Crust," *U.S. Geol. Surv. Prof. Paper 127* (1924); L. S. Selivanov, "Chlorine and Bromine in Massive Crystalline Rocks," *Compt. rend. (Doklady) Acad. Sci. U.R.S.S.*, Vol. XXVIII (1940), pp. 809–13; T. Fellenberg and G. Lunde, "Contribution à la géochemie de l'iode," *Norsk Geol. Tids.*, Vol. IX (1926), p. 48.

New determinations of fluorine in sea water have been made by T. G. Thompson, and H. J. Taylor, "Determination and Occurrence of Fluorides in Sea Water," *Indust. Eng. Chem., Anal. Ed.*, Vol. V (1933), pp. 87–89. See also V. M. Goldschmidt, "The Principles of the Distribution of the Chemical Elements in Minerals and Rocks," *Jour. Chem. Soc., London* (1937), pp. 655–73.

A large surplus of both chlorine and bromine is present in the sea. In sharp contrast to this, only a fraction of 1 per cent of the fluorine contained in primary rocks reaches the sea and remains in solution. By far the greatest part is returned to the lithosphere. In which rocks does it occur, and in which minerals is it concentrated? Partial answers to these questions may be obtained from the following data.

FLUORINE IN THE OSLO REGION

The region around Oslo, Norway, forms a well-known and well-defined petrographic province, containing deep-seated rocks as well as lavas and progressively metamorphosed sediments. The deep-seated rocks form a series of mildly alkaline character, ranging from monzonites through syenites and quartz-syenites to alkali-granites with a nepheline-syenite offshoot. The petrographic features of the series are in part apparent from Figure 1. A half-hundred fluorine analyses of these rocks have been worked out.[8] Of interest to this paper are the results from the igneous rocks. In Table 2 the average contents of fluorine and phosphorus pentoxide of each rock type have been listed.

Table 2 shows that the phosphorus content is not related to that of fluorine in a regular way, thus again supporting Shepherd's conclusion that apatite is not the chief carrier of fluorine. This is particularly true for the later members of the magma series—nordmarkites, ekerites, and granites, which are virtually without apatite; in them most of the fluorine is fixed in biotite and hornblende (see Fig. 2).

On the other hand, attention should be directed to three individual analyses of kauaiite (so-called "Oslo-essexite"), a coarse-grained basalt or gabbro, occurring in necks or pipes that are supposed to represent feeding channels for the extensive basalt flows of the province. In these basalts we find that the proportion

[8] T. F. W. Barth and B. Bruun, "Fluorine in the Oslo Petrographic Province," *Norske Vid. Akad. Skr. No. 8* (1945), pp. 5–12.

between fluorine and phosphorus is very nearly 1:10, indicating that all fluorine is fixed in apatite.

FLUORINE IN ROCKS AND WATERS FROM ICELAND

Determinations of fluorine and phosphorus in unaltered Icelandic basalts have been listed in Table 3. The amount of fluorine is, in each case, about one-tenth that of phosphorus, which, checking with the percentage of apatite determined in thin sections, indicates that all the fluorine is fixed in apatite. Since basalts of the Oslo region show the same relation, it would seem a natural conclusion that, generally, fluorine in basalts (and probably in gabbros) is locked up in apatite.

Two samples of dike rock from northern Iceland, strongly altered (by acid thermal waters?), have been collected by L. Hawkes and analyzed by E. S. Shepherd,[9] who found them high in fluorine; the samples gave 0.156 and 0.338 per cent fluorine.

Magmatic emanations are manifestly high in fluorine, and studies of the thermal activity in Iceland have demonstrated[10] that the acid hot springs and fumaroles are more closely related to the magmatic emanations than is the water of the alkaline hot springs, which is partly meteoric and has reacted with rocks.

From areas of acid hot springs I have only one observation: Hverarönd, east of Myvatn in northern Iceland, is an area of intense solfataric activity. The acid hot springs and fumaroles have boiled to pieces and "dissolved" basaltic rocks, eventually transforming them into clay. Basalts in various stages of alteration can be found. For analysis a basaltic lava was sampled from a crack or vent through which an intense vapor exhalation was taking place. The pores of the lava were full of incrustations, mostly of

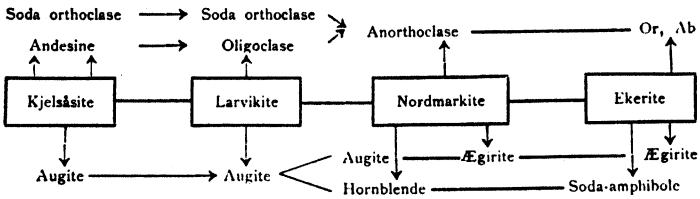

FIG. 1.—Differentiation diagram of the Oslo plutonic rocks. Lardalite is a nepheline syenite that branches off from larvikite. Akerite is a structural variety chemically related to kjelsasite.

TABLE 2

AVERAGE CONCENTRATION OF FLUORINE AND OF PHOSPHORUS IN THE PRINCIPAL TYPES OF THE OSLO IGNEOUS ROCKS

No.	Rocks	Per Cent F	Per Cent P_2O_5
1	Akerite	0.070	0.51
2	Kjelsåsite	.090	.52
3	Larvikite	.078	.59
4	Lardalite	.090	.62
5	Nordmarkite	.086	.20
6	Nordm-Ekerite	.072	.08
7	Ekerite	.067	.04
8	Granite	.063	.07
9	Kauaiite	.09	.91
10		.05	.55
11		0.07	0.73
Weighted average		0.075	0.305

[9] "The Gases in Rocks and Some Related Problems," *Amer. Jour. Sci.*, Vol. XXXVA (1938), pp. 311–51.

[10] T. F. W. Barth, "Pristine and Contaminated Rock Magma and Hot-Spring Water," *Bull. volcanologique*, Ser. II, Vol. VI (1940), pp. 83–87.

sulphates (the full analysis of the rocks is given in Table 4). As seen from the analysis, this lava obviously has received fluorine delivered by the acid gases.

Alkaline hot-spring water is remarkably low in fluorine; the concentration is in the same range as that encountered in ordinary river water. Silica sinter deposited by hot springs is also low in fluorine (see Table 5). Since the hot-spring water is fed by magmatic emanations, one would expect it to contain much fluorine, unless fluorine was extracted by interaction with the contiguous rocks.

TABLE 3

UNALTERED BASALTS FROM ICELAND

Locality	Per Cent F	Per Cent P_2O_5	P_2O_5/F
Ellidaárvogur	0.002	0.02	10.0
Svinahlid	.006	.08	13.3
Hveradal	.020	.21	10.5
Ash from Vatnajökull	.022	.23	10.5
Littli Geysir	0.040	0.42	10.5
Average	0.018	0.192	10.6

In order to test this, two analyses were made.

1. A pebble of basaltic composition was found in the crater of a small, boiling spring close to Geysir. For a long time the pebble has been exposed to the boiling water (pH=8.5); it was obviously altered by the water and should have had a good chance to absorb or assimilate fluorine, if there were any. The analysis shows, however, that no fluorine was assimilated; on the contrary, the ratio P_2O_5/F indicates that some fluorine was lost.

2. A drill core of basaltic tuff was taken at 200 meters depth at Reykir in Mosfellssveit, an area of exceptional thermal activity, where the underground is soaked with hot-spring waters (pH=8.0). Again the analysis demonstrates that no fluorine was deposited in, or assimilated by, the tuff[11] (see Table 5).

The analyses of Tables 4 and 5 indicate, therefore, that at least part of the fluorine of the primary magmatic emanations is contained in the acid hot springs

TABLE 4

ALTERED LAVA EAST OF HVERARÖND

	1		2
	Weight Per Cent	Atom Per Cent	Weight Per Cent
SiO_2	42.21	49.6	10.61
TiO_2	1.87	1.6	0.47
Al_2O_3	7.69	10.6	7.84
Fe_2O_3	11.87	10.6	3.55
FeO	1.70	1.7	5.70
MnO	0.02	.....	.....
MgO	0.96	1.7	1.53
CaO	5.40	6.8	2.72
Na_2O	1.80	4.1	0.22
K_2O	0.52	0.8	0.07
H_2O^+	6.09	(45.2)	5.79
H_2O^-	5.42		24.66
CO_2	0.22	0.4	.....
P_2O_5	0.20	0.2	.....
SO_3	13.52	11.9	37.17
F	0.48	(1.8)	.....
Cl	0.02	.....	.....
Total	100.01	100.0	100.33

NOTES TO TABLE 4

None: ZrO_2, Cr_2O_3, BaO, rare earths.

1. Analysis of altered lava east of Hverarönd, northern Iceland (Barth analyst).

2. Analysis of alunogen efflorescences, Námafjall, northern Iceland (Bruun analyst).

CALCULATED MINERAL COMPOSITION OF ALTERED LAVA

Quartz and opal	31.4
Albitic feldspar	26.5
Hematite, ilmenite	12.8
Pyroxene	2.0
Calcite, apatite, fluorite	2.1
Gypsum	8.0
Alunogen*	17.2
	100.0

* Composition as stated in col. 2.

[11] A rhyolite from Yellowstone, altered by alkaline thermal waters, was tested by Fenner and found to contain no fluorine (see in E. T. Allen and A. L. Day, "Hot Springs of the Yellowstone National Park," *Carnegie Inst. Wash. Pub. No. 466* [1935], p. 471).

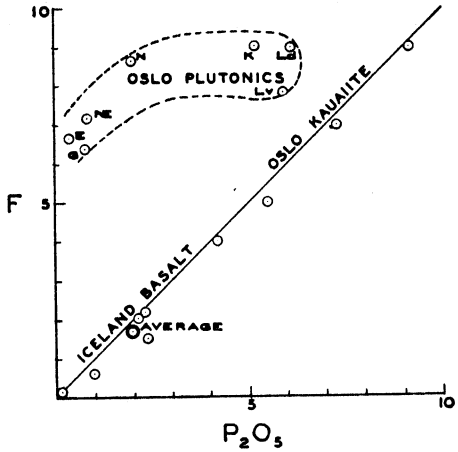

FIG. 2.—Concentration of fluorine and phosphorus in various rocks. *Ordinate:* content of F in hundredths of per cent; *abscissa:* content of P_2O_5 in tenths of 1 per cent. *E* = ekerite, *G* = granite, *K* = kjelsasite, *Ld* = lardalite, *Lv* = larvikite.

TABLE 5

CONCENTRATION OF FLUORINE IN HOT SPRINGS AND IN ROCKS ALTERED BY ALKALINE HOT-SPRING ACTION

Name of Spring	pH	F (Mg. per Liter)	Rock	Per Cent F	Per Cent P_2O_5	P_2O_5/F
Spytir	8.0	1.45	Silica sinter*	0.003	Nil	.....
Reykir in Olfus	8.0	0.55	Basaltic pebble†	.015	0.24	16
Fosshver	8.2	Nil	Drill core‡	0.006	0.06	10

* From Stekkjatunshver. † From crater of hot spring. ‡ From Reykir in Mosfellssveit.

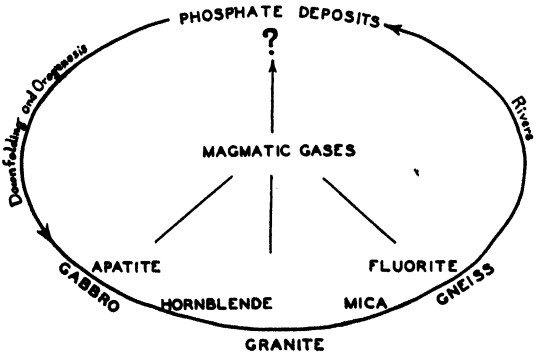

FIG. 3.—The geochemical cycle of fluorine

and fumaroles but that the alkaline hot springs have lost all primary fluorine before they emerge at the surface. But the mechanism of this extraction of fluorine is not known, nor is the location where it takes place.

CONCLUSION

It would seem that the cycle of fluorine is something like this: From inland waters and from the sea fluorine is extracted by phosphates.[12] Thereby, mineral deposits with a composition resembling fluorapatite are formed. These deposits represent a station on the road back to the igneous rocks.

The close relation of phosphorus to fluorine, so important in the sedimentary cycle, does not persist through subsequent transformations to other types of rock. In the lithosphere, fluorine and phosphorus gradually become separated as is evidenced by the fact that in the late-magmatic rocks and in the high-metamorphic sediments fluorine occurs in such minerals as mica and hornblende that contain no phosphorus.

Figure 3 illustrates graphically the geochemical cycle of fluorine.

[12] G. R. Mansfield ("The Role of Fluorine in Phosphate Deposition," *Amer. Jour. Sci.*, Vol. CCXXXVIII [1940], pp. 863–79) thinks that fluorine is caught by phosphorus compounds and locked up in insoluble deposits, whereas fluorine compounds without phosphorus generally are dissolved and carried away. Since phosphates and fluorine seem to have great affinity for each other, any conditions affecting phosphate deposition might tend also to localize the trapping of fluorine and account for its spotty occurrence in sea-bottom deposits, as noted by Shepherd.

TECTONIC FEATURES OF THE UTLEY META-RHYOLITE

O. E. GRAM

ABSTRACT

In south-central Wisconsin, outcrops of pre-Cambrian igneous rocks occur as monadnocks and ledges surrounded by Paleozoic sediments of Cambrian and Ordovician age. A careful study of such features as flow structures, chilled zones, inclusions, spheroids, gash veins, and an apparent sill leads to the conclusion that there is a minimum of eight separate flows, which were later subjected to stress in a horizontal shearing movement, which resulted in the development of an overturned anticline and syncline. During the field seasons of 1938 and 1939 the writer made a detailed study of the Utley meta-rhyolite, about 17 miles south of Berlin, Wisconsin, and a reconnaissance survey of the other igneous rocks in the area. A map of the quarry at Utley was made by a pace-and-compass survey, no other maps being available.

IGNEOUS ROCKS IN THE AREA

GENERAL RELATIONS

The Utley meta-rhyolite is one of a group of pre-Cambrian igneous rocks which occur at ten separate localities in south-central Wisconsin in an area of about 750 square miles (Fig. 1). Six of the outcrops are acidic extrusives, classed as devitrified rhyolites or meta-rhyolites; three are intrusive granites; and one is a close-to-the-surface, or hypabyssal, intrusive rhyolite. W. C. Alden[1] reports that the presence of other pre-Cambrian monadnocks beneath the Paleozoic sediments is shown by well-borings at several places in the area. In nearly all the localities the acid rocks are cut by basic dikes of altered diabase or greenstone, and in a few cases the main igneous body and the basic dikes are cut by dikes of granite porphyry.

The relation of the igneous rocks to the Paleozoic sediments is one of nonconformity. A conglomerate marks the base of the upper Cambrian St. Croix sandstone wherever it is in contact with the underlying igneous masses, the included pebbles and boulders being rhyolite fragments derived from the weathering of the pre-Cambrian rocks before Cambrian submergence. The St. Croix is the only Paleozoic sediment in direct contact with the pre-Cambrian masses.

CHEMICAL COMPOSITION AND AGE OF THE ROCKS

Chemically, the igneous rocks exposed in the area show a fairly uniform composition, and this similarity is the basis for assuming a common age. This chemical similarity extends beyond the area and into the Baraboo region, where the rhyolites have been described by J. T. Stark[2] as pre–Middle Huronian in age. Thus the Utley meta-rhyolite and the other igneous masses in the area are assumed to be pre–Middle Huronian in age.

Complete analyses of these rocks were made by W. W. Daniells, of the University of Wisconsin, and published by W. H. Hobbs and C. K. Leith.[3] Table 1 shows these analyses, with an analysis of the Baraboo rhyolite included for comparative purposes.

As shown by the table, the rocks of this area are all characterized by a high silica content. Another prominent fea-

[1] "The Quaternary Geology of Southeastern Wisconsin," *U.S. Geol. Surv. Prof. Paper 106* (1918), p. 51.

[2] "Igneous Rocks in the Baraboo District, Wisconsin," *Jour. Geol.*, Vol. XL, No. 2 (1932), pp. 119–39.

[3] "Pre-Cambrian Volcanic and Intrusive Rocks of the Fox River Valley, Wisconsin," *Univ. Wis. Bull. 158, Sci. Ser.*, Vol. III, No. 6 (1907), pp. 259–63.

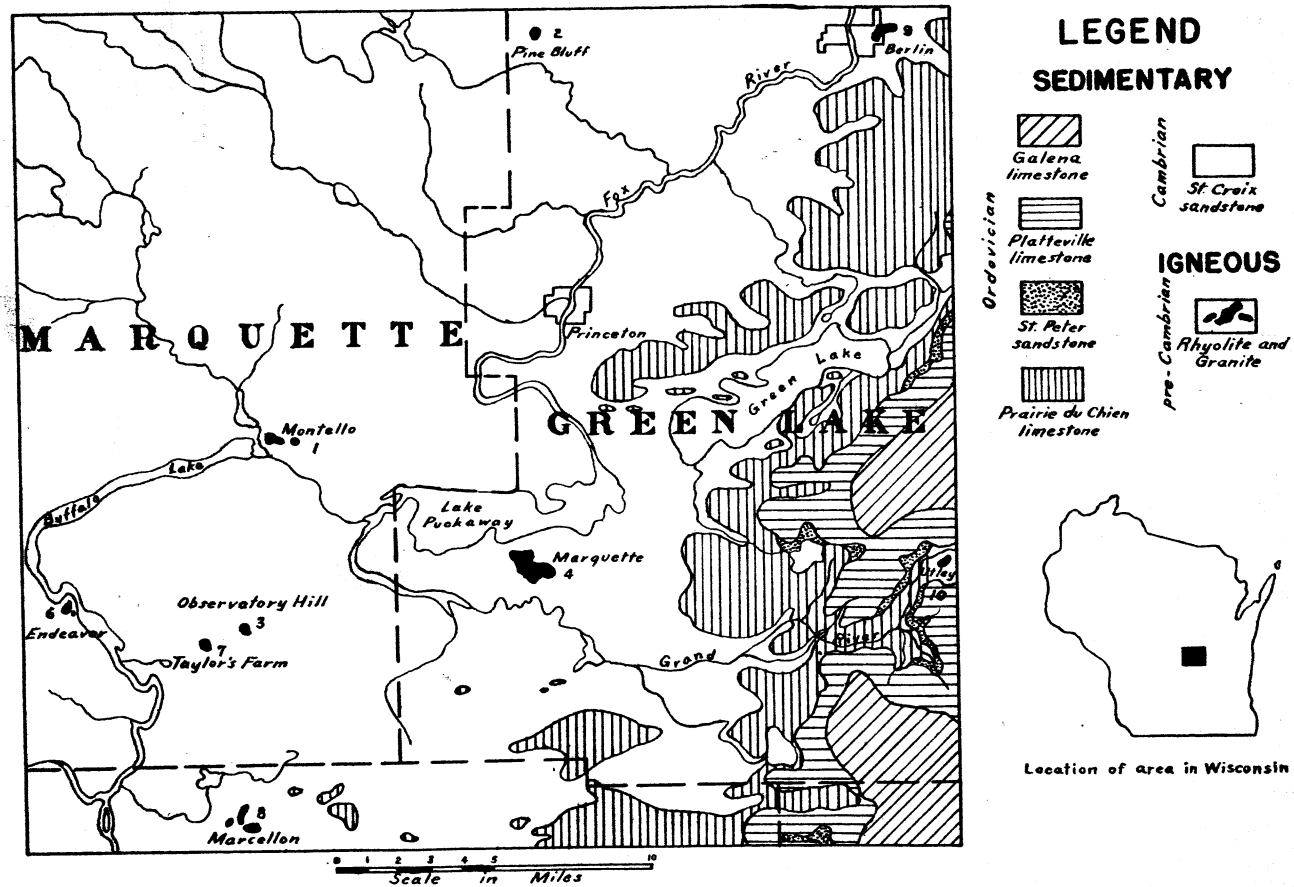

FIG. 1.—Geologic map showing location of igneous rock outcrops

ture is the predominance of soda over potash in many of the localities. In general, these chemical relationships are so striking that they strongly suggest a consanguinity of the rocks.

PREVIOUS WORK IN THE AREA

The most comprehensive work done in the area was that of Samuel Weidman[4] in 1898. He described three of the outcrops, including Utley, in detail, dealing mainly with the microscopic petrography of the rocks.

Another fairly detailed study of the area was made in 1906 by W. H. Hobbs and C. K. Leith.[5] They discussed the surface volcanic features in detail and cited the lithologic and chemical similarities of the rocks as proof of their consanguinity.

Since Hobbs and Leith published their report, no further detailed geologic work has been carried on in this area, although quarrying operations continuously exposed new surfaces until 1929.

[4] "Pre-Cambrian Igneous Rocks of the Fox River Valley, Wisconsin," *Wis. Geol. and Nat. Hist. Surv., Bull. 3* (1898).

[5] Pp. 247–78 of ftn. 3.

GENERAL GEOLOGY OF THE UTLEY META-RHYOLITE

PETROGRAPHIC AND CHEMICAL CHARACTER

Technically the rock at Utley should be referred to as a "porphyrytic rhyolite" in order to signify an extrusive rock as distinct from an intrusive, which, with the same petrographic character, would be termed a "rhyolite porphyry." In addition, microscopic examination plainly shows that the rhyolite has been metamorphosed, and, in conformity with the usage of the United States Geological Survey, it should be called a "meta-rhyolite." Strictly speaking, therefore, the rock at Utley is a porphyritic meta-rhyolite. For convenience in the following discussion this technical nomenclature will be disregarded, and the rock will be called a "meta-rhyolite."

The characteristic texture of the Utley meta-rhyolite is porphyritic. Quartz and feldspar phenocrysts are present in approximately equal proportions. The quartz phenocrysts possess good crystal form, have a limpid appearance, and are fairly uniform in size, averaging about $\frac{1}{16}$ inch in diameter. The feldspar phenocrysts are flesh-colored to white, are very

TABLE 1

CHEMICAL ANALYSES OF THE IGNEOUS ROCKS

	1*	2	3	4	5	6	7	8	9	10
SiO_2	73.78	74.62	74.46	73.30	71.24	72.80	78.23	79.03	73.65	73.09
Al_2O_3	17.18	10.01	15.28	15.32	12.20	15.50	11.11	13.23	11.19	13.43
Fe_2O_3	.....	3.83	1.95	1.21	1.71	2.04	1.73	0.34	1.31	2.57
FeO	1.64	1.72	0.74	0.96	5.44	0.60	1.03	0.18	3.25	.....
CaO	0.85	2.43	0.92	1.33	0.98	0.52	0.28	0.25	2.78	2.29
MgO	.....	0.33	0.08	0.39	0.13	0.08	.....	0.07	0.51	1.03
K_2O	4.48	3.38	3.01	3.86	1.86	2.52	4.08	2.28	1.86	1.58
Na_2O	2.82	3.33	2.57	3.47	4.29	5.70	3.44	3.95	3.74	3.85
H_2O	0.12	0.24	0.58	0.26	0.81	0.43	0.25	0.19	0.44	0.72

* 1, Montello granite; 2, Pine Bluff and Waushara granites; 3, Observatory Hill rhyolite; 4, Marquette rhyolite; 5, Baraboo rhyolite; 6, Endeavor rhyolite; 7, Taylor's Farm rhyolite; 8, Marcellan rhyolite; 9, Berlin rhyolite; 10, Utley meta-rhyolite.

conspicuous by their color contrast with the dark groundmass, and show a wide variation in size, ranging from less than $\frac{1}{16}$ to $\frac{3}{8}$ inch in diameter. The groundmass is uniformly aphanitic and almost glassy in the hand specimen, although microscopic examination reveals a porphyritic texture. In color the groundmass is very dark to black, a somewhat uncommon color in rhyolites.

Quartz and feldspar are the only minerals recognizable megascopically; but the type of feldspar cannot be determined, although the flesh color is suggestive of an alkalic variety. Microscopic examination of thin sections reveals two varieties of feldspar which occur either alone or together in an interlocking structure known as "microperthite." The more common variety has the extinction angle of albite, and it is probable that the lesser variety is microcline. No orthoclase was found in any of the thin sections.

Under the microscope the groundmass of the meta-rhyolite shows a definite porphyritic texture, with quartz and feldspar making up most of the larger crystals. The presence of hornblende in the original rock is shown by altered phenocrysts with the characteristic crystal outline of hornblende. In all cases several new minerals occupy the space of the original hornblende. Samuel Weidman[6] lists eight such minerals, namely, biotite, sericite, quartz, magnetite, epidote, sphene, apatite, and zircon. The writer noted all these minerals except sphene and zircon.

The smaller crystals in the groundmass are more difficult to identify. Here Weidman lists eleven minerals, including biotite, muscovite, sericite, calcite, hornblende, sphene, epidote, apatite, zircon, magnetite, and pyrite. In addition to these minerals, the writer noted a considerable amount of chlorite, relatively small amounts of kaolinite, fairly large amounts of calcite, and two other minerals which are probably iddingsite and limonite or goethite. Fluorite is present as a secondary infiltration product, apparently having been deposited from solutions circulating along fissures. Siderite is common as a secondary filling of vugs and cavities in the spheroid zones.

[6] Pp. 30–31 of ftn. 4.

PROOF OF DEVITRIFICATION

Although the groundmass of the meta-rhyolite is holocrystalline in its present state, evidence exists to prove that this crystallization is of secondary nature and that the groundmass was originally a glass. This devitrification is in keeping with the fact that few, if any, pre-Cambrian of Paleozoic rhyolites with glassy bases have ever been reported, for all of them have been more or less completely devitrified.

The primary evidence of an initially glassy character of the groundmass is the presence of perlitic cracks, spherulites, and crystallites. According to F. F. Grout[7] and many other petrographers, these textures develop only in glasses, and their presence in a holocrystalline rock is taken as a sign of an original glassy texture.

Further evidence that the Utley meta-rhyolite is a devitrified glass is the disappearance under crossed nicols of some of the original textures of the rock. Perlitic and fluxion textures are easily recognizable in ordinary light, but these textures are lost under crossed nicols in a homogeneous holocrystalline groundmass. In the few instances in which perlitic partings do not disappear, there is pronounced crystallization along the partings. Spherulites are rare under

[7] *Petrography and Petrology* (New York: McGraw-Hill Book Co., Inc., 1932), p. 115.

crossed nicols and always tend to grade imperceptibly into the crystalline groundmass.

Devitrification also is suggested by the mottled texture common in the meta-rhyolite in which crystals of quartz and feldspar include small patches of minerals identical with the groundmass. These patches have apparently resulted from devitrification of glass which occupied embayments in quartz and feldspar crystals, a feature frequently observed in the phenocrysts of rhyolites and attributed to resorption. The mottled texture is not truly poikilitic, but because of its close resemblance might be called "pseudo-poikilitic."

The presence of spherulites is a common occurrence in a glass, as spherulites represent rapid growths in a quickly cooling magma. The relative scarcity of spherulites in the Utley meta-rhyolite suggests that they have been obliterated in the process of devitrification, and the gradation of those that are present into the holocrystalline groundmass suggests that they are partially obliterated remnants of the originals.

STRUCTURAL GEOLOGY OF THE UTLEY META-RHYOLITE

EXTRUSIVE CHARACTER OF THE ROCK

Field evidence suggesting that the Utley meta-rhyolite is an extrusive body is by no means abundant, but the few structual features present are fairly conclusive of such an origin. Flow structure is evident in one locality and is suggested in others. Chilled zones are present. Oval-shaped inclusions are distributed throughout the mass of rhyolite, and their elongation in a common direction suggests orientation in a flow. Paralleling the flow structure are several layers of spheroids, which suggest weathered surfaces of separate flows. Gash veins, developed by flows slipping over one another during folding, and volcanic breccia are other features supporting the extrusive nature of the rhyolite (Fig. 2).

The field evidence is strongly supported by laboratory evidence. In thin sections many textures may be seen which are characteristic of flows. The devitrified groundmass indicates an original glass, which in itself is suggestive of a flow. The presence of numerous crystallites, including margarites, globulites, and single and radiating trichites, gives evidence of the rapid cooling of the original magma. Fluxion or flow texture is prominent and can be observed in most of the thin sections. Generally, this texture is accentuated by the presence of numerous magnetite grains arranged in lines parallel to the direction of flow. Devitrification causes this texture to disappear under crossed nicols. Perlitic partings are common, and they are suggestive of rapid cooling, with consequent contraction in an extrusive body.

STRUCTURAL FEATURES

THE LAYERS OF SPHEROIDS

Occurring in several different zones throughout the mass of meta-rhyolite at Utley are bodies of ellipsoidal shape caused by weathering along spheroidal partings. In describing these bodies, Weidman[8] appropriately used the term "spheroid." The spheroids vary in size from $\frac{1}{4}$ to 2 or more inches in diameter. Generally, they are elongated in a northeast-southwest direction. They are composed of a core having a porphyritic texture and containing the same minerals as the normal meta-rhyolite (but with a greater proportion of kaolinite and sericite) surrounded by an alteration zone made up of kaolin, sericite, and chlorite. The kaolinite and sericite suggest weath-

[8] Pp. 6–7 of ftn. 4.

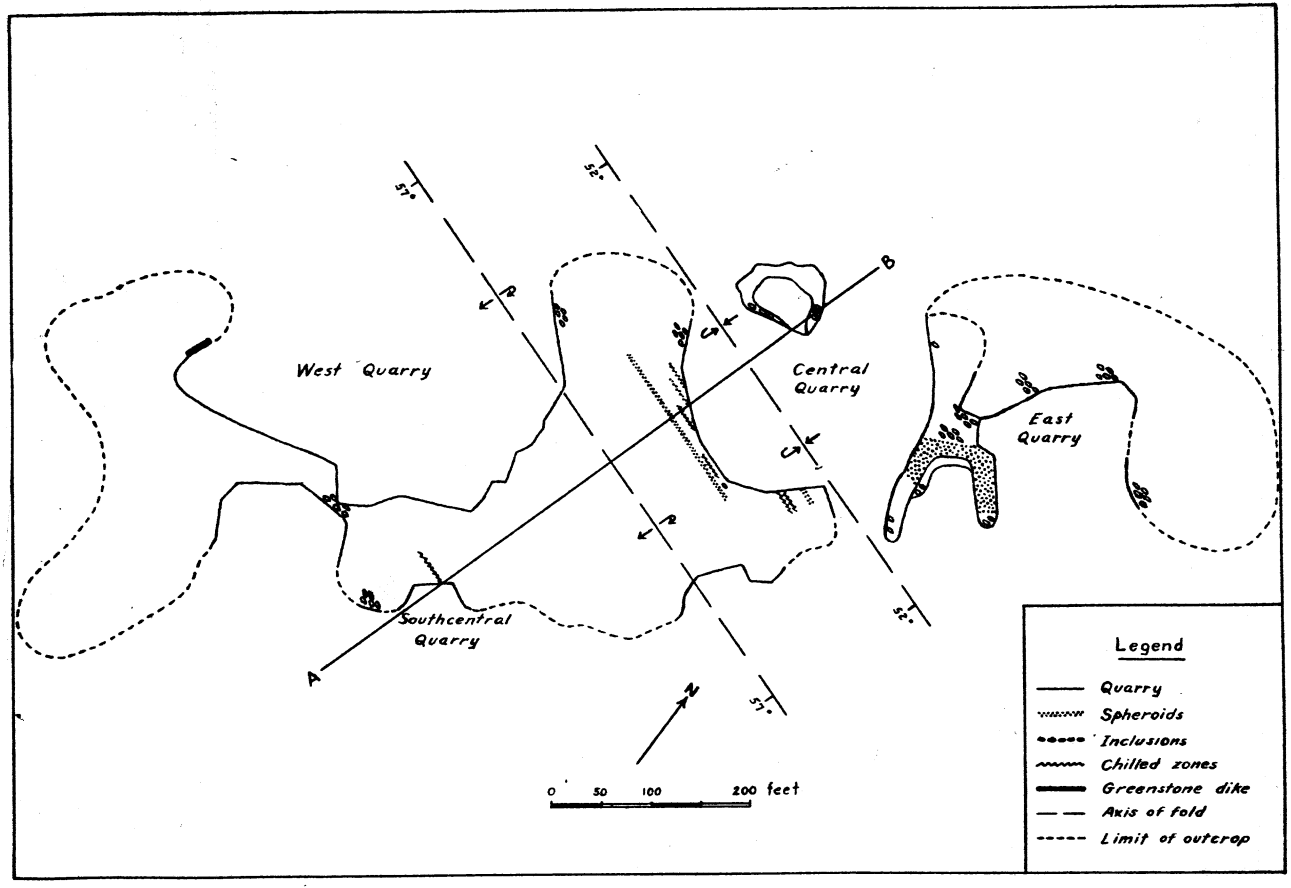

FIG. 2.—Structure map of the Utley quarry

ering of feldspars, which may have formed during devitrification along the spheroidal partings.

The spheroidal zones apparently represent tops of flows, for the most clearly defined zone shows a sharp contact with the meta-rhyolite to the north and a gradation into that to the south. If this interpretation is correct, it is probable that water solutions had easy access to the spheroid zones, with a resultant weathering of minerals along partings.

Many of the spheroids contain cavities or vugs partially or entirely filled with secondary minerals. Quartz, in well-formed crystals, is most common. Other minerals are siderite, limonite, calcite, kaolinite, and hematite. The origin of these cavities is not entirely clear, but escaping gases probably played a major role in their formation.

The spheroids occur in at least five different zones (Fig. 2), two of which probably represent the top of the same flow. One zone follows the present surface of the meta-rhyolite in the northeastern part of the outcrop. This zone is characterized by large numbers of nutlike masses or nodules on a weathered surface. The thickness of the zone is from 1 to 2 feet, but part of it may have been eroded or planed off by glacial action. Dipping 20° to the south, this zone probably represents a continuation of one of the other four spheroidal zones.

The other four layers of spheroids occur in the outcrop south and southeast of the central quarry. All dip between 80° and 85° to the south and strike N. 70° W. to N. 80° W. Their thicknesses vary from a few inches to more than 4 feet. The thickest zone occurs on a vertical face in the southeast part of the central quarry and also on the floor of the quarry west of this face. The spheroids in this zone are well defined, but more or less widely scattered, in comparison with the closely packed spheroids of the other zones. About 35 feet south of this zone is the most clearly defined layer of the outcrop, about 15 inches wide and made up of closely packed spheroids. The spheroids in this zone show a sharp contact with the meta-rhyolite to the north and a gradation into the meta-rhyolite to the south.

Between the two layers of spheroids described above is a zone on the southwest side of the central quarry which includes at least two more layers. The spheroids in these layers are small, varying between $\frac{1}{4}$ and $\frac{1}{2}$ inch in diameter, and on a weathered surface they give the appearance of rings and chains. These layers are about 4 feet apart, and both dip 82° S. and strike N. 70° W. They can be traced for only a few feet, for to the east the meta-rhyolite has been removed by quarrying, and to the west they disappear into the vertical face of the quarry wall.

THE VOLCANIC BRECCIA

On the south face of the central quarry is a zone in which angular fragments of meta-rhyolite occur in a dark felsitic groundmass. These fragments show a small amuunt of resorption around their outer boundaries, this resorption being best shown by a slight reaction rim around the feldspar crystals and by embayments in quartz crystals. The long axes of most of these fragments parallel the strike of the spheroid layers and the other structural features. The boundaries of this zone are indistinct because of the quarrying of the meta-rhyolite to the north and the weathering of the rock to the south.

In discussing this zone, Weidman[9] implies that it is not a volcanic breccia be-

[9] *Ibid.*

cause there is no trace of vesicular texture. However, a volcanic breccia is not necessarily characterized by such a texture. Many flows fail to exhibit a vesicular texture either because there was a minimum of gas in the magma or because the viscosity of the magma was so high that the gas could not escape.

According to the definitions of various igneous breccias as given by Grout,[10] these angular fragments constitute an eruptive breccia, in contrast to flow breccia and contact breccia. In all probability the fragments represent previously solidified rhyolite in the walls of the conduit which were shattered and dispersed in a flow. This flow differs from others in the area in its extremely fine-grained felsitic character, which probably was due to a relatively high viscosity of the magma. Microscopic examination of thin sections reveals a high silica content, which undoubtedly increased the viscosity. The relatively small amount of resorption of the fragments and the concentration of the breccia in a small area are also suggestive of a magma of high viscosity, for in a more fluid magma there would be expected more resorption of the included fragments, as well as a wider dispersion of them throughout the flow.

FLOW STRUCTURE

A few feet northwest of the breccia is a limited zone in which excellent megascopic flow structure is developed. Striking N. 85° W. and dipping 82° S., this structure is clearly defined by weathering, which causes small ridges to stand out on the surface, where the crystals have a linear arrangement. The strike reading cannot be considered accurate because of the very limited exposure of the structure.

[10] P. 119 of ftn. 7.

INCLUSIONS

Widely scattered throughout all parts of the outcrop are many incusions which are ellipsoidal masses of meta-rhyolite, differing from the normal rock in the greater size of the quartz and feldspar phenocrysts and in the extreme darkness of the groundmass. They vary in size from 1 to 8 inches in their long diameter, and most of them are elongated in an east-northeast direction. Many occur singly, but others are found in groups (Fig. 2.).

Dip readings were obtained in only a few instances, for most of the inclusions are seen on horizontal surfaces. Single inclusions on vertical faces of the east quarry vary in dip from 18° to 22° S., with the exception of one, which dips 7° S. On the northeast wall of the pit in the central quarry, ten inclusions in two groups about 12 feet apart occur on a vertical face, and all of them dip 22° S. An interesting characteristic is the flatness of these inclusions in the plane of dip and strike, all being three to six times as long in the direction of dip as they are normal to it. This suggests flattening or deformation through folding.

The origin of the inclusions is not entirely clear. They could have originated in at least three different ways, namely, (1) fragmental material blown into the air in a semiplastic state and falling back into the flows, (2) erosional fragments of solidified flows being engulfed in fresh flows, and (3) solidified rhyolite, which was torn loose from the walls of the conduit by rising magma, which flowed out onto the surface. The ellipsoidal shape of the inclusions can be explained by any of the three theories; but the scattered distribution, the large size of the phenocrysts, and the character of the groundmass are most satisfactorily explained by the third theory. If this hypothesis is

correct, the inclusions were formed in the same manner as the eruptive breccia; but, because they were included in a flow of lower viscosity, their angularity was destroyed by resorption, and they assumed an ellipsoidal shape oriented in the direction of the flow. The increased size of the phenocrysts is probably due to renewed crystal growth when the fragments were included within the hot magma, and the increased size of grain in the groundmass is similarly explained. Reaction rims around phenocrysts near the contact between the inclusions and the meta-rhyolite are common, as would be expected under this theory.

GASH VEINS

On the east face of the central quarry is a series of small quartz stringers and lenses, which appear to be gash veins, filling fractures which presumably developed during folding of the flows. In one zone the veins run out horizontally to the south from two joint planes, which dip 18° S. In the other zone, a few feet to the north, the veins run out horizontally on both sides of a joint plane. A theoretical application of horizontal compression and vertical tension to these zones suggests that the rhyolite above each joint moved north with respect to the meta-rhyolite below, this conclusion being in harmony with the structure of the outcrop.

CHILLED ZONES

Near the top of a vertical face in the south-central quarry is a zone of dark aphanitic rock about ½ inch thick, striking N. 80° W. and dipping 30° S. Presumably representing the chilled surface of a flow, it is the only clue to the structure of the southern part of the outcrop.

In the central part of the outcrop, on the southeast side of the central quarry, are two well-defined chilled zones, which represent the top and bottom of a flow, a sill, or the edges of a dike. The relations of this body to the surrounding meta-rhyolite are confusing. The contact between the chilled zones to the south and the meta-rhyolite is very irregular, being fairly smooth near the base of the quarry, quite wavy halfway to the top, and extremely jagged near the top. The other chilled zone is more regular. Between these two zones the rock grades into a meta-rhyolite that is almost identical with the main mass of meta-rhyolite. In a few places the chilled zone disappears, and the porphyritic meta-rhyolite runs out to the contact, the latter, however, remaining distinct. In the meta-rhyolite on both sides of this body is a well-defined banding which is due to an alignment of feldspar phenocrysts parallel to the contact. This banding is noticeable only in the lower part of the zone.

The strike of this body cannot be determined with any degree of accuracy because of its very limited exposure at the top of the quarry, but the small zone exposed strikes approximately N. 65° W. and dips 82° S. to vertical. In thickness it ranges from 2 feet near the base of the quarry to almost 5 feet at the top. Mineralogically its composition is identical with the main mass of meta-rhyolite.

No conclusive evidence exists to prove whether this body is a flow, a dike, or a sill, but the writer favors a sill. The fact that the banding of feldspar phenocrysts occurs on both sides of the body suggests that it is an intrusion rather than a flow, for exomorphic effects of a flow would be limited to the rhyolite beneath the flow. As previously stated, the coarse grained phase of this body in places extends all the way to the contact, thus eliminating the chilled zone. This is an improbable occurrence in a flow but is not uncommon in an intrusive body. Further evidence

favoring an intrusion is negative and includes a lack of fluxion texture in thin sections, a lack of xenoliths, and a lack of vesicular texture.

INTERPRETATION OF THE STRUCTURE

From the foregoing evidence there seems little doubt that the Utley meta-rhyolite is a series of extrusions. Eight separate flows are recognized, although in some cases the evidence is meager. Spheroids mark the upper surface of at least four, and possibly five, flows. The sill-like body with its chilled zones presumably separates two flows, for if it is a sill it would be expected to intrude between flows. The chilled zone in the south-central part of the outcrop marks the top of another flow. The volcanic breccia may or may not mark a separate flow, and the spheroid zone between the central and east quarries may represent the surface of still another flow unless it is a continuation of one of the spheroid layers to the southwest. The thickness of the definitely established flows ranges from 4 to 40 feet or more.

The attitude of the various features described in the preceding pages gives a definite clue to the geologic structure of the meta-rhyolite outcrop. All features with sufficient exposure to permit accurate strike readings to be taken show a strike ranging from N. 70° W. to N. 80° W. and the remaining features with limited exposures closely approximate this strike. The dip readings obtained are strongly indicative of an overturned anticline (Fig. 3) with a south flank dipping 30° S., as shown by the chilled zone in the south-central quarry, and a north flank overturned to 82° S., as shown by the sill (?), chilled zones, and spheroids on the south side of the central quarry. The spheroids and inclusions in the northern part of the outcrop show a reversal, forming a syncline with a north flank dipping approximately 22° S. The axes of the folds strike N. 75° W., and the axial plane of the overturned anticline dips 57° S., while that of the syncline dips 52° S. (Fig. 2).

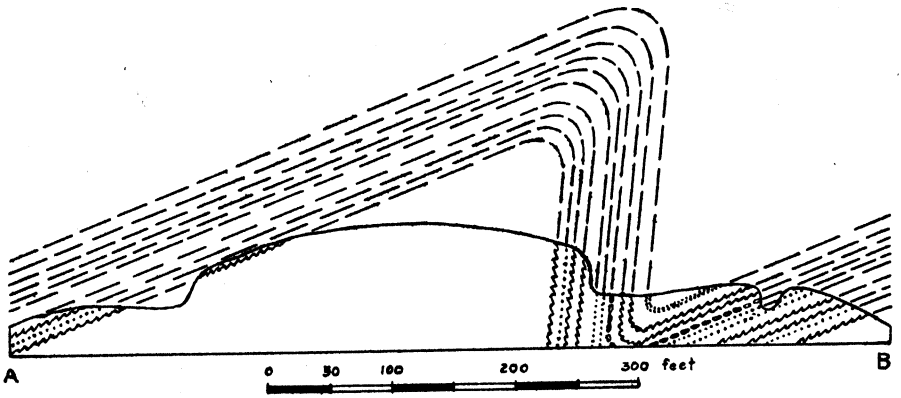

FIG. 3.—Theoretical structure section along line A–B. (See Fig. 2 for legend)

THE JOINT SYSTEM

In an attempt to determine the type of regional stress which caused the deformation of the rhyolite flows at Utley, the writer spent several days mapping the major joints or master-joints. After eliminating from consideration those joints

which were apparently related to the joints in the associated sediments in the area, three prominent sets were discovered, these sets striking N. 70° E., N. 2° E., and N. 32° W. Other joints were grouped into less prominent sets, the most prominent of which had a strike of N. 70° W., which parallels the strike of the folds, thus suggesting that they are tension joints. This conclusion is further justified by the very steep dip of these joints, their curved and irregular surfaces, and the fact that their occurrence is limited mainly to the area which roughly corresponds to the crest of the anticline.

Both the steeply dipping N. 32° W. and the N. 70° E. joints show characteristics of true shear joints in their smooth, plane surfaces. Nearly horizontal slickensides are common, especially along the planes of the N. 70° E. set. If these are true shear joints, they should furnish a clue to the direction of regional stresses.

Resistance to shearing along fracture planes is a composite of the strength of the rocks against shearing and their frictional resistance to motion.[11] The rupture has been proved theoretically to be 45° with the direction of stress, provided that friction is neglected. Actually, the effect of friction will cause this angle to approach 30°–35° with the direction of compressive stress in brittle material. According to the shear theory, the direction of maximum compressive stress will always bisect the acute angle between planes of fracture, provided that the stress is nonrotational. This theory cannot be applied to the Utley meta-rhyolite because the direction of maximum stress would have had to be from N. 71° W., in order to bisect the acute angle between the major shear planes. This direction parallels the axes of folding and is therefore impossible. G. F. Becker,[12] however, suggested that fracturing occurs where the strain is greatest. In other words, the plane of maximum strain will be a plane of fracture, and the obtuse angle between sets of fractures will face the direction of nonrotational stress.[13] By applying this theory, known as the "strain theory," to the joints and folds at Utley, the direction of maximum compression would have been from N. 19° E. A nonrotational stress from such a direction would develop the existing joint pattern and the folds at Utley almost perfectly, for the stress would be at right angles to the strike of the folds, would bisect the obtuse angle between the shear joints, and would produce the N. 70° W. tension joints. The N. 2° E. joints would not be entirely in accordance with this pattern, but friction could cause the discrepancy from the theoretically perfect development. However, an application of the strain ellipsoid to this pattern shows that under a nonrotational stress of this type it would be necessary to have the easiest relief lateral instead of vertical, in order to obtain the joint pattern existing at Utley. Because it is difficult to reconcile easy lateral relief with folding of any great extent and because compressive stresses are more often applied tangentially with a resulting rotational strain, a horizontal nonrotational stress such as described above must be considered to be in a doubtful category.

If a rotational stress in the form of a horizontal shearing movement is applied to the Utley rhyolite, the existing joint

[11] C. M. Nevin, *Principles of Structural Geology* (New York: John Wiley & Sons, Inc., 1936), p. 17.

[12] "Finite Homogeneous Strain, Flow, and Rupture of Rocks," *Bull. Geol. Soc. Amer.*, Vol. IV (1893), pp. 13–90.

[13] Recent laboratory experiments by Griggs have tended to discredit Becker's analysis.

pattern and the folds can be explained. Experimental work by W. J. Mead[14] showed that horizontal shearing movements first produced tension cracks inclined about 45° to the direction of movement. These fractures were immediately followed by two sets of nearly vertical shear fractures, one striking parallel to the direction of movement and the other parallel to the free side. Folding in such a case developed at approximately right angles to the first tension fractures. If the deformation at Utley was due to a horizontal shearing couple from N. 32° W. and S. 32° E., the first fractures to form would tend approximately N. 13° E. Since friction would cause a variation in the inclination of these fractures to the direction of shearing movement, the N. 2° E. joints might easily be the primary tension fractures. The two sets of shear fractures would then correspond to the N. 70° E. joints, which closely parallel the free side, and the N. 32° W. joints, which parallel the direction of movement. At the same time, folding would occur, and the axes of the folds would trend N. 77° W., which closely corresponds to the trend of the folds at Utley and which would give rise to tension fractures, such as those striking N. 70° W. However, the folds developed in Mead's experiments are difficult to explain because of the N. 71° W. easy lateral relief developed under a rotational stress of this type.

The writer feels that no definite conclusions regarding stress conditions in the area can be drawn until such time as all the other igneous outcrops are studied in detail. The joint system at Berlin, about 17 miles north of Utley, is extremely complicated and does not fit the pattern at Utley. However, the Berlin rhyolite offers one clue in its very pronounced cleavage system, this cleavage paralleling the northeast axis of the outcrop. Since cleavage develops in a plane normal to the shortening, it is probable that the stress at Berlin operated from a northwesternly or southeasterly direction. This fact suggests that a horizontal shearing movement from N. 32° W to S. 32° E. at Utley must be considered as a strong possibility.

ACKNOWLEDGMENTS.—The writer is especially indebted to Dr. J. T. Stark, of Northwestern University, for his assistance in the field and laboratory and for his valuable suggestions and criticisms; to Dr. A. L. Howland, who supplied several valuable suggestions pertaining to the petrographic character of the rhyolite and who obligingly checked the writer's microscopic identification of minerals in thin sections of the rhyolite; and to Dr. R. M. Garrels for his field assistance as well as his stimulating suggestions pertaining to some of the theoretical data. The writer also acknowledges the co-operation of the Wisconsin Granite Company, which granted permission to work in the quarry.

[14] "Notes on the Mechanics of Geologic Structure," *Jour. Geol.* Vol. XXVIII (1920), pp. 505–23.

DISCOVERY OF LETTERS BY LYELL AND DARWIN

HENRY P. ZUIDEMA
University of Michigan

ABSTRACT

Original letters written by Sir Charles Lyell to the Rev. Charles Kingsley and the Rev. John Pye Smith and also a letter from Charles Darwin to James Croll on alternate glaciation of northern and southern hemispheres were found recently between the pages of the first edition (Vol. III) of Lyell's *Principles of Geology*. The Lyell letters do not appear in previously published biographical material on Lyell. In his letter to Kingsley, Lyell expresses his interest in the exploration of the Green and Colorado rivers by John Wesley Powell. To Dr. Smith he reveals his belief that the waves of immigration of Europeans into the United States in the 1840's, "like the irruptions of the Goths & Vandals, are doing their utmost to overwhelm the indigenous civilization, but against which a noble stand is making."

While reading Volume III of the first edition of the *Principles of Geology* by Charles Lyell[1] recently, the writer was agreeably surprised to discover two original letters written by Lyell and, on further examination of the book, a letter sent by Charles Darwin to James Croll on November 24, 1868, on the subject of alternate glaciation in the northern and southern hemispheres. The letters, loosely placed between the pages of the *Principles*, are in excellent condition. The Detroit bookseller from whom the writer purchased the book, with the companion volumes, was unable to provide any clue to their previous ownership, and the writer has no pertinent information which would permit tracing the letters along their evidently devious travels.

The Lyell letters are to the Rev. Charles Kingsley, written April 26, 1871, and to the Rev. John Pye Smith (June 2, 1849). From an examination of the known Lyell correspondence, these appear to be heretofore unpublished letters. The letter from Darwin to Croll evidently was written by Mrs. Darwin but is signed by Darwin and is mentioned in the two-volume compilation, *More Letters of Charles Darwin*.[2]

Few records of Lyell's activity during the six years prior to his death in 1875 are preserved, but these few show that the great uniformitarian retained his keen interest in geology to the end and showed no evidence of declining mental strength.

The note to Kingsley is on stationery bearing the Lyell address, 73 Harley Street, London W, embossed in blue, and was written four years before Lyell's death, as he was revising his tenth edition of the *Principles*. The letter reads as follows:

MY DEAR KINGSLEY

On my return from a geological Easter tour to Devon and Cornwall, I find your very kind note & very curious & instructive (?) Report by Major Powell on the cañons of the Rio Colorado. I must read it again when the work is out as I cannot yet fully understand the country. The *chines* as they call them in Hants near Xchurch & Bournemouth are I suppose diminutive cañons.

I hope soon to hear an account of yourself and family.

My wife desires her kindest remembrances & believe me

Ever most truly ys
CHA LYELL

"The 'chines' of Hants (Hampshire) and the Isle of Wight are, in the words of Horace B. Woodward, "gullies which

[1] *Principles of Geology* (3 vols.; London: John Murray, 1830–33).

[2] *More Letters of Charles Darwin*, ed. Francis Darwin and A. C. Seward (London: John Murray, 1903).

73, Harley St
London. W.

April 26
1871

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pour dissimulation cañons & hope soon to hear an account of yourself & family. My wife desires her kindest remembrance & believe me

Yrs
Cha Lyell

have been formed by the action of springs in making their way over the cliffs into the sea."[3]

Lyell's excursions frequently took him to the coast in the Christchurch-Bournemouth area, where the Eocene strata have been deeply incised.

When the letter was written, Major Powell was at Green River, Wyoming, preparing for his second trip down the Green and the Colorado and had probably already made his observation that "the Uintas rose no faster than the Green could erode."

The energetic and versatile Powell, who had lost part of an arm at Shiloh and had collected fossils while his men dug trenches before Vicksburg, now was destined to write one of the most dramatic chapters in American geology. He could not have known that the vast canyons of the Southwest were being compared across the Atlantic to "the chines in Hants" or that Lyell, the master whose ideas Powell had championed in America in the face of stubborn opposition, was eagerly awaiting publication of Powell's papers, "the better to understand the country."

The four tours of Lyell in the United States three decades earlier had taken him only to the Mississippi. We can only wonder what impetus he might have given to the development of American geology, had circumstances permitted him to travel into the then still untamed West.

Kingsley, canon of Chester, sociologist, and geologist, is better known as the author of *Westward Ho!* and *The Water Babies*. He also wrote *Town Geology*, an introduction to geology for the people of Chester. He met Darwin in 1854, at the Linnaean Society and recalled Darwin as "a quiet, meek man, very anxious to know whether I really 'denied the atonement,' on which point, I think, I satisfied him."

Lyell met Kingsley in 1860, the year in which the exuberant Huxley, slapping the knees of a friend next to him and uttering, "The Lord hath delivered him into my hands," launched into his reply to the Bishop of Oxford, who was at the time attacking Darwin's newly published *Origin of Species*.

Darwin, Lyell, and Kingsley kept up a sprightly correspondence through the years, and the influence of the great naturalists on the theologian-geologist is apparent in Kingsley's letter to a friend, the Rev. F. D. Maurice, in 1863:

> I am very busy working out points of Natural Theology, by the strange light of Huxley, Darwin and Lyell. The state of the scientific mind is most curious; Darwin is conquering everywhere, and rushing in like a flood, by the mere force of truth and fact. The one or two who hold out are forced to try all sorts of subterfuges as to fact, or else by evoking the *odium theologicum*.

The four visits of Lyell to the United States were made during a great tide of immigration, such as the country had never seen before. In 1844 alone, 297,000 European emigrants came to America. In 1845, during his second visit, 114,000 were admitted. The Irish famine brought another deluge of newcomers in the succeeding years, and by 1850 the total population was in excess of 23,000,000.

What he saw in Boston, in New York, and in the rude hinterland and the babble of foreign tongues that he heard caused Lyell much concern, as witness his letter to Dr. Pye Smith in 1849, following his return to England:

> MY DEAR SIR
>
> I have given orders to my publisher to send a copy of my book to Jackson & Wolfe 18 St. Paul's Ch y'd, of which I beg your acceptance. It is entitled a Second Visit to the United States & I hope that none of my allusions in the 11th

[3] *Geology of England and Wales* (London: Longmans, Green & Co., 1876).

and 12 Chapters v. 1. to the Congregational or Independent Churches are offensive to your views in reference to the bearing of the organization of such churches on the educational movement now in progress in those parts of the Union which are least kept back by emigration from Europe which like the irruptions of the Goths & Vandals are doing their utmost to overwhelm the indigenous civilization but against which a noble stand is making.

Believe me
Sincerely yours
CHA LYELL

Dr. Pye Smith

The Rev. John Pye Smith (1774–1851) was pastor for forty-three years of the (Independent) Gravel Pits Chapel at Homerton, a Fellow of the Royal Society, and author of *The Mosaic Account of the Creation and the Deluge Illustrated by the Discoveries of Modern Science* (1837) and *Scripture and Geology* (1839).

The library of Dr. Smith no doubt already held a copy of Lyell's *Travels in North America, with Geological Observations on the United States, Canada and Nova Scotia*,[4] the nongeologic chapters therein containing this paragraph:

The state of Ohio has always punctually discharged the interest of her debt by direct taxes imposed for that special purpose, although there has been a deficit from the beginning on the proceeds of her public works. She is of recent origin, and her growth has been more rank and luxuriant than that of any other State of the Union. An influx of illiterate Irish, Welsh and Westphalian settlers, has tended to lower the educational qualifications of her electors, considered as a whole; but she came of a good New-England stock, which, like the philosopher's stone, has converted much of baser metal into gold.

Now, four years after the appearance of this volume, Dr. Pye Smith, upon receiving his complimentary copy of *Second Visit to the United States*, found additional fears concerning the future of the Union in Lyell's description of a "repeal meeting" in Boston, where "an orator with an Irish accent, addressed the crowd on the sufferings of the Irish people precisely as if he had forgotten on which side of the Atlantic he then was!"

The Darwin letter of 1868 to James Croll, one of the documents found between the pages of my copy of Lyell's *Principles*, is as follows:

DEAR SIR

I have read with the greatest interest the last paper which you have kindly sent me. If we are to admit that all the scored rocks throughout the more level parts of the United States result from true glacier action it is a most wonderful conclusion & you certainly make out a very strong case; so I suppose I must give up one more cherished belief. But my object in writing is to trespass on your kindness & ask a question, which I dare say I c'd answer for myself by reading more carefully, as I hope hereafter to do, all your papers; but I shall feel much more confidence in a brief reply from you.

Am I right in supposing that you believe that the glacial periods have always occurred alternately in the northern and southern hemispheres; so that the erratic deposits which I have described in the S. parts of America & the glacial work in New Zealand c'd not have been simultaneous with our glacial period. From the glacial deposits occurring all round the Northern hemisphere & from such deposits appearing in S. America to be as recent as in the North, & lastly, from there being some evidence of the lower descent of glaciers all along the Cordillera, I inferred that the whole world was at this period cooler. It did not appear to me justifiable without distinct evidence to suppose that the N. and S. glacial deposits belonged to distinct epochs; tho' it wd have been an immense relief to my mind if I cd have assumed that this had been the case.

Secondly, do you believe that during the glacial period in one hemisphere, the opposite hemisphere actually becomes warmer, or does it merely retain the same temperature as before?

I do not ask these questions out of mere curiosity, but I have to prepare a new ed. of my Origin of Species, & am anxious to say a few words on this subject on yr authority. I hope you will excuse my troubling you.

Pray believe me
Very faithfully yours
(*Sgd.*) CHARLES DARWIN

Darwin was seeking an explanation for the observed distribution of plants

[4] 2 vols.; London: John Murray, 1845.

11 Harley St
June 2d /49

My dear Sir

I have given orders to my publisher to send a copy of my book to Jackson & Walford 18 St. Pauls Ch. Yd. of which I beg your acceptance. It is entitled a second Visit to the United States & I hope that none of my other sons in the 11th & 12 chapters v. 1. to the congregational or Independent Churches are opposed to your views in reference to the bearing of the organization of such churches on the educational movement in progress in those parts of the Union which are least kept back by emigration from Europe which like the irruption of the Goths & Vandals are doing them almost to overwhelm the indigenous civilization but against which a noble stand is making.

Believe me
sincerely yrs
Chas Lyell

Dr Pye Smith

I do not ask these questions out of mere curiosity; but I have to prepare a new ed. of my Origin of species, & am anxious to say a few words on this subject on yr authority: I hope that you will excuse my troubling you.

Pray believe me
very faithfully yours

Charles Darwin

Bern
Bromley
Kent. S.E.

Nov 24 1868

Dear Sir

I have read with the greatest interest the last paper which you have kindly sent me. If we are to admit that all the scored rocks throughout the more level parts of the United States result from true glacier action it is a most wonderful conclusion, & you certainly make out a very strong case; so I suppose I must give up one more cherished belief. But my object in writing is to trespass on your kindness & ask a question, which I dare say I'd answer for myself by reading more carefully, as I hope hereafter to do, all your papers; but I shall feel much more confidence in a brief reply from you. Am I right in supposing that you believe that the glacial periods have always occurred alternately in the northern & southern hemispheres; so that the ~~glacial~~ erratic deposits which I have described in the S. parts of America, & the glacial work on New Zealand i'd not have been simultaneous with our glacial period. From the glacial deposits occurring all round the Northern hemisphere, & from such deposits appearing in S. America to be as recent as in the North, & lastly, from there being some evidence of the former lower descent of glaciers all along the Cordillera, I inferred that the whole world was at this period cooler. It did not appear to me justifiable without distinct evidence to suppose that the N. & S glacial deposits belonged to distinct epochs, tho' it wd have been an immense relief to my mind if I'd have assumed that this had been the case. Secondly, do you believe that during the glacial period in one hemisphere, the opposite hemisphere is actually ~~rendered~~ becomes warmer, or does it merely retain the same temperature as before?

and animals and was turning more and more to the astronomical theory of Croll (1821–90), the Scottish geologist and climatologist, and away from the opinion of his old friend, Lyell, who consistently maintained that geographical changes are the principal, and not the subsidiary, causes of glaciation.

A zealous disciple of Lyell ever since he had first read Lyell's *Principles* as a youth while on the epochal cruise of the "Beagle," Darwin now chose to differ from Lyell regarding the validity of Croll's views. In 1866, two years before the letter to Croll was written, Lyell wrote to Darwin as follows:

> I have been doing my best to do justice to the astronomical causes of former changes of climate, as I think you will see in my new edition [the tenth, first vol. published in November, 1866, the second in 1868] but I am more than ever convinced that the geographical changes are, as I have always maintained, the principal and not the subsidiary ones. . . . . In my winter of the great year, I gave you in 1830 cold enough to annihilate every living thing. The ice now prevailing at both poles is owing to an abnormal excess of land, as I shall show by calculation. Variations in eccentricity [of the orbit] have no doubt intensified the cold when certain geographical combinations favored them, but only in exceptional cases, such as ought to have occurred very rarely, as paleontology proves to be the case.
>
> Ever most truly yours,
>
> CHARLES LYELL

For Darwin, however, the Croll theory gained in appeal, and his letter to Croll followed. This letter was followed by another from Darwin, a few months later, in which Darwin wrote Croll:

> Sir C. Lyell, who is staying here, is very unwilling to admit the greater warmth of the S. hemisphere during the Glacial period in the N; but, as I have told him, this conclusion which you have arrived at from physical considerations, explains so well whole classes of facts in distribution, that I must joyfully accept it; indeed, I go so far as to think that your conclusion is strengthened by the facts of distribution. . . . .

In 1872, Darwin, in his sixth edition of the *Origin of Species*, abandoned his cautiously worded statement of earlier editions, in which he indicated belief in a world entirely colder during glacial times. He now gave considerable space to Croll's ideas, adding that "this [Croll's] conclusion throws so much light on geographical distribution [of animals and plants] that I am strongly inclined to trust it. . . . ."

In the same year there appeared Lyell's eleventh edition of the *Principles of Geology*, revised, and Lyell stands firm, stating:

> Mr. Darwin in the last edition of his Origin of Species has inclined toward adopting Mr. Croll's theory of alternate glaciation and perpetual spring in the opposite hemispheres, on the ground that it would account for some of the anomalies in the distribution of animals and plants, by affording a refuge for tropical life during a period of extreme cold.
>
> But . . . . until we know what climate the countries now inhabited by tropical animals and plants were enjoying in glacial times it is premature to contend with imaginary difficulties as to the survival of forms which would have been extinguished if the snow and ice had been universal down to Lat. 55, even over one hemisphere at a time.

Today, after seventy-five years of research and speculation, Lyell's view of the problem still agrees with that of leading students of the Pleistocene, a tribute to the sagacity of the man to whom Darwin dedicated the second edition of his journal[5] of the voyage of "H.M.S. Beagle" with his acknowledgment that "the chief part of whatever scientific merit this journal and the other works of the author may possess, has been derived from studying the well-known and admirable *Principles of Geology*."

[5] Charles Darwin: *Journal of Researches into the Natural History and Geology of the Countries Visited during the Voyage of H.M.S. Beagle Round the World* (London: John Murray, 1852).

REVIEWS

The Pleistocene Period: Its Climate, Chronology and Faunal Successions. By FREDERICK E. ZEUNER. London: Ray Society; sold by Bernard Quaritch, Ltd., 1945. Pp. xii+322; figs. 76. 46*s*.

This book is "entirely concerned with the evidence for the relative and absolute chronology of the Pleistocene, for the various types of environment existing at certain times in certain areas, and with the evolution of the fauna in deposits of known age" (p. iii). Data on climatic and geographic changes and on stratigraphy are reviewed and marshaled into a relative, stratigraphic chronology of the Ice Age; the modern astronomic hypothesis of the Pleistocene glacials and interglacials is reviewed; and the relative and the absolute chronologies are merged. The evolution of the modern mammals receives special attention. The treatment is essentially confined to the Old World.

An introductory chapter is devoted to some principles of Pleistocene stratigraphy (pp. 1–29). The subdivisions of this age are primarily based on changes in climate (p. 2); and the evidence is supplied by till, moraines, and other glacial deposits, loess, frost phenomena, soils due to chemical weathering, river terraces, fauna, and flora. Because an advancing ice normally destroys older glacial deposits, the formations of the periglacial belt furnish more complete records (p. 28). These foundations of the relative chronology are lucidly though briefly discussed.

Chapters ii–iv deal with the relative chronology proper (pp. 30–135). In northern Germany three separate glaciations are distinguished, viz., the Elster, Saale, and the Last Glacial, corresponding to the Mindel, Riss, and Würm (p. 38). The Last Glacial probably included the Warthe, Weichsel, and Pomeranian stages. However, the age of the Warthe, which is comparable to the Iowan of North America, remains an unsolved problem (pp. 33, 35, 39). In the Alps there may have been four main glaciations, as Penck and Brückner established in 1909, using glaciofluvial gravels: the Günz, Mindel, Riss, and Würm (p. 41). Late studies are believed to prove that the Würm comprised three maxima; each of the others, two maxima (pp. 47, 48); but this is not conceded by Albrecht Penck.[1] Besides, there appear to have been several pre-Günz glacial ages. An attempt at correlating the major Pleistocene climatic ages in Scandinavia-Germany, the Alps, and North America is presented on page 53.

The periglacial zone is the belt of frost climate surrounding an ice sheet (p. 55). While the modern periglacial zone coincides with the tundra, that of the glacials included also a peripheral loess steppe belt which graded into the forest zone. The loess steppe was dry and cold. Its subsoil was permanently frozen, and its highest monthly mean was probably below 10° C. (50° F.). The periglacial area, which was temperate during interglacials and interstadials, has furnished the most complete record of the climatic variations during the Pleistocene (p. 56). Climatic fluctuations are recorded by river terraces within the zone (pp. 25, 56). During the oncoming of a glacial the forests withdrew from high summits and ridges and frost action increased, supplying streams with an abundance of debris, which they carried downvalley. When the glaciation waned, the vegetation grew denser and returned to higher levels, and the streams began to sink their beds in the deposits of the preceding cold age, converting them into terraces. The river terraces reveal at least eight cold ages, grouped in four pairs (p. 63). Climatic changes are also recorded by a succession of loess beds (cold and dry climate) and weathering soils (temperate and humid climate) (p. 64). In Germany there are at least two young loesses of the Last Glacial and three old loesses (p. 73). A correlation of periglacial river terraces and loess beds with glacial stages is suggested on page 79.

The climatic chronology of northern France, based on loess and solifluction recording the glacials and on weathering indicating the interglacials and interstadials, has been correlated with low (glacial) and high (interglacial) sea-levels, respectively, which controlled the terraces of the Somme River (pp. 92, 99).

To sum up: In continental Europe the Last

[1] "Die Strahlungstheorie und die geologische Zeitrechnung," *Zeitschr. d. Gesellsch. f. Erdkunde zu Berlin*, Jahrgang 1938, pp. 321–50; see p. 336.

Glacial comprised two main stages and regionally a third stage. The two intermediate glacials each had two stages, and the earliest glacial two stages in the Alps, one stage elsewhere. The interglacials were mild (p. 99). This climatic succession, based purely on geological observations, is called the detailed relative chronology (p. 100).

Geographical separation, oceanic climate, and several centers of glaciation make it difficult to establish a Pleistocene climatic chronology for the British Isles (p. 101). However, the till deposits seem to indicate three cold ages of the Last Glacial, two extensive glaciations which probably were correlatives of the Saale and the Elster, and two minor cold ages in Crag times which probably corresponded to the Günz (p. 114). In the Thames basin buried channels, fillings, and abandoned terraces record glacials, interglacials, and interstadials which seem to correspond to similar stages on the Continent (pp. 133, 135).

Chapter v, pages 136–65, is entitled "The Astronomincal Theory as the Basis of an Absolute Chronology of the Pleistocene." The astronomical elements considered are the inclination of the earth's axis (the obliquity of the ecliptic; mean length of periodicity forty thousand years), the eccentricity of the earth's orbit (mean length of periodicity ninety-two thousand years), and the longitude of the perihelion (precession of the equinoxes; mean length of periodicity twenty-one thousand years). Formulas for calculating the secular changes of these elements were developed by the American astronomic mathematician J. N. Stockwell in 1873, and the calculations were made by the German, Ludwig Pilgrim, in 1904. From Pilgrim's determinations the Serbian mathematician and physicist Milutin Milankovitch in 1920 deduced the consequent secular fluctuations of the solar radiation received at the outer limit of the earth's atmosphere; and in 1924, at the request of the German climatologist Wladimir Köppen, he expressed the radiation changes on the latitudes 55°, 60°, and 65° N. during the summer half-year of the last six hundred and fifty thousand years in the form of latitudinal equivalents. Köppen interpreted Milankovitch's radiation graph as a graph of the Pleistocene glacials and interglacials, and he correlated the glacials with the minima of summer radiation and the associated maxima of winter insolation.

Later the Serbian astronomer V. Michkovitch, using the formulas of 1843 of the French mathematician U. J. J. Leverrier, duplicated Pilgrim's calculations of the secular changes of the mentioned astronomical elements; and Milankovitch again calculated the corresponding radiation changes and constructed a graph of the summerly variations on latitude 65° N. (1930). The good agreement between this graph and that of 1924 based on Stockwell's formulas is an important check. In 1930 and 1938 Milankovitch presented the calculated secular fluctuations of the radiation during the winter as well as during the summer half-year on latitudes 5°, 15°, 25°, 35°, 45°, 55°, 65°, and 75° N. and S. for the last six hundred thousand years (p. 146). While the amount of the annual solar radiation has fluctuated little through the ages, those of the caloric summer and winter half-years have varied considerably (pp. 145–49). Low summer solar radiation is accompanied by high winter radiation, and high summer insolation by low winter radiation.

As stated, the equable oceanic ages of low summer and high winter radiation were already associated by Köppen in 1924 with the ages of ice expansion. This correlation was confirmed by the German climatologist Walter Wundt, who found that a reduction of the annual temperature range by 1° C. depresses the snow line by an amount of 78 meters (p. 152), and by Milankovitch, who showed that the height of the modern snow line in the different latitudes closely correlates with the summer radiation (pp. 150–51). With cooler summers the melting of snow and ice decreased, and the snow line fell; and with milder winters the snowfall increased because of its being greatest near the freezing temperature, and the zone of maximum snowfall rose in altitude. This latter rise, together with the greater descent of the snow line, made the two levels approach each other or coincide, augmenting the snow available for glaciers (pp. 152–54). The magnitude of the climatic effects of the calculated fluctuations of the solar radiation is difficult to assess, but Zeuner modestly assumes that they lowered the snow line by 200–400 meters (pp. 152, 156). Of course, mountains are a primary condition for the development of glaciers in middle and low latitudes.

Thus the mountains and the primary effects of the low summer and high winter radiation were insufficient to produce the large Pleistocene ice sheets. However, the glaciers and ice caps that did form further modified the climate in various ways. First and foremost, they caused an increase in the percentage of the solar radi-

ation which was reflected, for the albedo of snow and ice is about 80 per cent, while that of fields, pastures, forests, and lakes is, at most, 10 per cent. This self-increasing capacity of ice sheets has been especially investigated and stressed by Wundt, who has calculated that the increased reflection during the maxima of the Pleistocene glaciations caused a temperature lowering on the Northern Hemisphere of over 4° C. (p. 156). This corresponds to a depression of the snow line of 600 meters, which is to be added to the original depression of a few hundred meters.

The ice sheets also brought about great changes in the air circulation: Glacial anticyclones formed and induced anticyclonic east winds, and the traveling cyclones were forced to take more southerly courses (pp. 156–58).

Zeuner concludes that the Pleistocene glaciations and associated climatic conditions in Europe are adequately explained by the principal summer minima and winter maxima of the solar radiation and by the supplementary effects of glaciers and ice sheets as they developed (p. 159). However, the Ice Age is not explained; for, although similar summer minima and winter maxima of insolation must have occurred periodically throughout the ages, the next preceding continental glaciation occurred in the Permian, two hundred million years ago (p. 161). The Pleistocene Ice Age apparently required some exceptional conditions favorable for glaciation; and these might have been, Zeuner believes, a Tertiary-Quaternary lowering of the sea-level, probably caused by sinking of the ocean floors and uplift of parts of the continents (p. 165).

Milankovitch's graph of the summer insolation agrees well with the geological sequences in Europe, especially those worked out by Penck and Brückner, Eberl, and Soergel (pp. 167, 170). The first extensive glaciation may have begun six hundred thousand years ago, and this date is taken by Zeuner as the demarcation line between the Tertiary and the Quaternary (p. 175).

Extension of the absolute astronomic chronology beyond the northern temperate zone meets difficulties, but Zeuner applies it for the Last Interglacial and the Last Glacial to the Mediterranean region (pp. 176–207).

Since the secular course of the summerly solar radiation is determined in high latitudes mainly by the inclination of the earth's axis, in low latitudes largely by the longitude of the perihelion and the eccentricity of the earth's orbit,[2] it changes gradually from a periodicity of forty thousand years near the poles to one of twenty-one thousand near the equator (pp. 149, 216). The main radiation minima and maxima of high latitudes occur on the entire Northern Hemisphere, but additional ones appear in mid-latitudes and grow in prominence toward the equator. It is unknown whether or not the radiation changes can explain the geologic and climatic history of the tropical zone, for their effects on the tropical climates have not been studied (pp. 217, 219). As a consequence, the Pleistocene history of tropical Africa can be correlated with that of Europe only on a general paleontological basis (pp. 208, 215).

Since the effects of the inclination of the earth's axis are identical on the Northern and the Southern Hemisphere, but those of the longitude of the perihelion and of the eccentricity are reversed, and since, as just stated, the relative influence of these factors varies with latitude, the secular course of the summer radiation in the two polar hemispheres was almost identical in high latitudes, almost reversed in low latitudes, and different in mid-latitudes.[3] In the latitudes of the main Pleistocene glaciations the culminations of the summer minima in the Northern and Southern hemispheres were offset in time at most ten thousand years, which would have made long-lasting glaciations largely contemporaneous. The climatic effects of the fluctuating summer radiation on the Southern Hemisphere are little known (p. 222).

Chapter ix, pages 225–52, deals with the Pleistocene fluctuations of the sea-level and the world-wide extension of the absolute chronology. Melting of the existing ice caps and glaciers would raise sea-level 40 to 60 meters, so that any higher abandoned shorelines must involve at least some vertical crustal movements (p. 225). In different parts of the world there are five high Quaternary shorelines or sea-levels between which there must have been relatively low ones (p. 247). The high sea-levels must be interglacial, the low ones glacial, as long held by glacialists (p. 248). The height of the old terraces decreases progressively from the oldest at 100 meters to the youngest at 7.5 meters (p. 246), a fact attributed by Zeuner to a continuous slow sinking of the sea-level (p. 248). The principal terraces and erosion channels are described, correlated, and dated (pp. 246, 252). The high sea-levels are particularly important

[2] Milutin Milankovitch, "Astronomische Mittel zur Erforschung der erdgeschichtlichen Klimate," *Handb. der Geophys.*, Band 9 (1938), pp. 593–698; see p. 661.

[3] *Ibid.*, pp. 660 and 661.

for a world-wide correlation and, Zeuner hopes, for extension of the astronomical chronology over the globe.

The concluding chapter, pages 253–78, treats the evolution of the mammals and of British land and fresh-water mollusks during the Quaternary. The composition of the faunas changed slowly by intrusion of newcomers and rapidly by a large-scale shifting with the environment (p. 256). The environment changed as the climate varied in respect to general temperature, seasonal contrast, and precipitation. In an oceanic climate the succession was from temperate forest through subarctic forest to tundra; in a continental climate from warm-continental steppe through loess steppe to tundra. However, the animals adjusted themselves very slowly, for faunas fully typical of the mentioned environments occur first in the Upper Pleistocene, that is, since the beginning of the Last Interglacial (p. 255). The most striking change in the mammal fauna was the increase of the cold component, which suggests a gradual adaptation to low temperatures (p. 274). This was brought about by migration and phylogenetic evolution. Several lineages of ascent are given. Since the earliest and the latest forms in a lineage are distinct species, it appears that half a million years were required for the evolution of a new species (pp. 277, 278).

Foremost among the many problems in this outstanding book is the absolute chronology of the Pleistocene. The modern hypothesis of the Pleistocene Ice Age employing the secular changes of the earth's position relative to the sun—which may be called the Milankovitch-Köppen-Wundt hypothesis—contains three main steps, viz.:

1. The Ice Age was made possible by a preglacial development of the most favorable conditions for glaciation since the Permian, and these conditions still prevail. The favorable changes included uplift of extensive and high mountain ranges in which glaciers could form, rise of continental shelves and submarine ridges so that these partly or practically excluded warm ocean currents from high latitudes, and lowering of the sea-level. The changes caused a temperature lowering.

2. Glaciers grew into ice fields and initial ice sheets, i.e., *extensive* glaciation was *started* by the climatic conditions induced by the extreme minima of the summer radiation and the associated maxima of the winter radiation, which were caused by certain, periodically recurring, positions of the earth in relation to the sun.

This was the most critical stage, for glaciers had to grow into effective reflectors of the solar heat, into small self-perpetuating ice sheets during the ten thousand to twelve thousand years that the periods of low summer and high winter insolation lasted. It thus depended on several factors and conditions whether or not the local glaciation developed into a continental glaciation. Some factors were: extent of glaciers and of sea ice at the onset of low summer insolation; the degree of change in the solar radiation; the time of beginning effectiveness of the expansion of the sea ice, of the reduction of the Gulf Stream water entering the Norwegian Sea, and of the reduction of the heat transport to high latitudes by the migratory cyclones; and the intensity of the summer radiation during the succeeding cycle. Even though the present-day ice recession suggests that glaciers are very sensitive to climatic change, a primary decrease of the annual temperature range by several degrees centigrade may have been necessary (cf. pp. 147, 149, 152). Thus, not every minimum of summer insolation, especially not the first one following upon an interglacial, need have been, or may have been, represented by ice sheets. For instance, there is no convincing evidence that ice sheets formed upon the radiation minimum 115,000 B.P. (before the present), though Zeuner so believes (pp. 53, 169). In America the Iowan glacial maximum, dated by G. F. Kay at 55,000 B.P., followed the radiation minimum of 72,000 B.P.; and the Mankato glacial culmination, as estimated at 25,000 B.P., approximately coincided with the radiation minimum of some twenty-three thousand years ago. Furthermore, since the double minima of summer radiation which are correlated with the first three large glaciations are spaced only forty thousand to forty-three thousand years, since the formation of ice sheets may have required more or less than twenty-five thousand years, and since there must have been a very considerable lag of the growth and shrinking of the ice sheets (pp. 160, 168), the continental glaciations and the sea-level fluctuations need not have been doubled.

3. The ice fields expanded into vast ice sheets with the help of secondary effects, especially cooling by the increasing reflection of the sun's heat (Wundt). This occurred at least partly during somewhat stronger summer insolation than is received today. The growing ice sheets caused the development of glacial anticyclones and anticyclonic winds and deflected the migratory cyclones. The pack ice expanded to fill the

North Atlantic southward perhaps to Cornwall and the banks off Nova Scotia.[4] Most drifting ice was carried westward on the north side of the mean storm tracks, but some reached the southerly belt of westerly winds and was transported by the anticyclonic whirl toward southwestern Europe and then south along the coast of Portugal. The Gulf Stream was deflected southward with the cyclone tracts until even its northeasternmost branches, now entering the Norwegian Sea, were incorporated in the clockwise whirl. The glacial anticyclones and the southerly position of the Icelandic and Aleutian barometric lows favored low-index circulation (weak westerlies in the middle latitudes,[5] during which strong cold waves penetrated exceptionally far south, but warm tropical air was carried only moderately far north. Therefore, while there is at present an enormous winterly importation of heat to Europe by the migratory cyclones and the Gulf Stream—the January temperature is on the Norwegian northwest coast 20° C. (36° F.), in central Europe 8° C. (14.4° F.) above the means for the latitudes—there was then little if any heat importation, and as a consequence the winter temperature may at the time of the glacial culminations have been low enough for the observed glaciations. Also the summer temperature, which was reduced by the ice sheets, by drifting ice and chilly waters of the Atlantic, and by fogs, may ultimately have become sufficiently low to account for the Pleistocene glaciers in Ireland, Portugal, and southwestern Yugoslavia which Penck[6] believed required a primary reduction of the annual temperature of at least 8° C. (14.4° F.).

These and other effects of the ice sheets assisted in spreading the glaciations and in creating the climatic conditions of the glacials.

The Milankovitch-Köppen-Wundt hypothesis of the Pleistocene Ice Age has a physical basis throughout and does not make unverifiable assumptions of, for instance, a lowered or fluctuating solar constant. Extensive mountain building and a general cooling preceded the Ice Age. Secular changes in the earth's solar position occur and have been calculated and dated for the last one million years; and they influence the earth's insolation and climate. The extreme minima of summer and maxima of winter insolation must have brought about at least a notable increase of glaciation; and the approximate coincidence of their dates and the estimated ages of the glacials suggests a causal relationship.[7] Secondary effects of ice sheets, some of them chain effects, are even at this early stage of their study known to have played an important part. Several independent as well as interrelated factors and causes rather than a few or a single one are in accord with the complexity of the climatic and geologic conditions and events of the Ice Age. For instance, widespread glaciation required primarily lower temperature in the British Isles and in the Scandinavian mountains and heavier snowfall in the Labrador Peninsula; the last great (Mankato, Tioga) ice advance between the Pacific and the Mississippi River coincided with ice retreat in the northeastern states. This composite hypothesis has changed in important respects and developed along new lines since first advanced by Köppen in 1924, and it is in the stage of growth. It opens a wide and promising field of research.

Zeuner's endeavors to correlate the geological history or relative chronology of the Pleistocene with the history of the fluctuating summer insolation therefore deserve serious and open-minded consideration. The probability that not every extreme minimum of summer insolation need have led to a continental glaciation and our inability as yet to appraise the several secondary effects of the ice sheets and the lag of expansion and of shrinking of the ice advise verification of details. Considerable study lies ahead.

Zeuner has rendered a great service by his able investigation and treatment of these exceptionally difficult fundamental problems of the Ice Age.

ERNST ANTEVS

Postglacial Forest Succession, Climate, and Chronology in the Pacific Northwest. By HENRY P. HANSEN. Philadelphia: Transactions American Philosophical Society, Vol. XXXVII (Part I), 1947. Pp. 130; figs. 114. $2.25.

[4] See also G. C. Simpson, "World Climate during the Quaternary Period," *Quart. Jour. Roy. Met. Soc.*, Vol. LX (1934), pp. 424–78; see pp. 431–37; *ibid.*, "Ice Ages," *Nature*, Vol. CXLI, No. 3570 (1938), pp. 591–98; *ibid.*, *Smithsonian Rept. for 1938*, pp. 289–302.

[5] C. G. Rossby, "The Scientific Basis of Modern Meteorology," *U.S. Dept. Agric. Yearbook 1941*, pp. 599–655; see pp. 621–31.

[6] Pp. 333–35 and 347–48 of ftn. 1.

[7] Albrecht Penck and Eduard Brückner, *Die Alpen im Eiszeitalter* (Leipzig, 1909), Vol. III, p. 1168; and G. F. Kay, "Classification and Duration of the Pleistocene Period," *Bull. Geol. Soc. Amer.*, Vol. XLII (1931), pp. 425–66.

Since Professor Hansen of Oregon State College began his western plant pollen studies in 1935, he has examined most peat bogs and many other pollen-bearing sediments in Oregon, Washington, and northern Idaho, has described the pollen profiles, and has discussed their significance in a long series of papers. Now he has brought together his pollen data from some seventy sedimentary columns in a single volume in which he treats the methodic, botanical, ecological, climatic, and geochronologic problems and draws important conclusions.

This review covers the chapters on geology, past climate, and geochronology (pp. 8–19, 33–35, 108–22). Other subjects in the monograph are: history and theory of pollen analysis, pollen-bearing profiles of the Pacific Northwest, rate of organic deposition, methods and technique, climate of the Pacific Northwest, forests of the Pacific Northwest, characteristics of Pacific Northwest forest trees, and postglacial vegetation history.

Most of the lake basins in which peat and other pollen-bearing deposits have accumulated in the Pacific Northwest were of direct glacial origin. Others were caused by damming of glacial streams, erosion by such streams, eustatic changes of sea level, and ponding of tributary streams by aggradation of the main valley. Still other basins were formed by landsliding, faulting, ponding by lava flows and beaver dams, and by severing of oxbows (pp. 8–19).

Layers of pumice and volcanic ash in the columns have served as excellent correlation markers (pp. 3, 37, 38, 114, 118, 119).

The culmination of the last continental glaciation, the Mankato, is given the usual age of 25,000 years, and the deposition of the pollen-bearing sediments resting on glacial drift is assumed to have begun some 18,000–20,000 years ago. The age of the pollen records has been arbitrarily called the Postglacial. On changes in the forest composition, shown for the individual regions in a table, this Postglacial has been divided into these four climatic periods (pp. 4, 35, 113–151:

Year	Period
0	the present
4,000	IV, cooler and moister
8,000	III, maximum warmth and dryness
15,000	II, increasing warmth and dryness
18,000–20,000	I, cool and moist

During the last glacial maximum the snow line in the Cascades was depressed some 1,400 meters (4,600 feet), according to Fritz Klute;[1] and the maxima of lodgepole pine and western white pine in the bottom layers of the sediments indicate cooler and moister climate than at any later time, but low percentage of pollen of boreal species suggests absence of tundra (p. 113).

A distinctly warmer age a few thousand years ago is recorded in several regions of the north temperate zone, and it has been dated in Sweden and Denmark at about 8000–4000 B.P. (before the present). The terminal date is also calculated from the historical salt contents in Owens Lake, California, and in Abert and Summer lakes, Oregon. This warm age was also dry in the western United States and is recorded not only by the forests of the Northwest but by deglaciation in the western mountains and by arroyo cutting and wind erosion in the Southwest (pp. 4, 111, 119).

A return to cooler and moister climate some 4,000 years ago is indicated in the western states by expansion of moisture-loving trees, rebirth of lakes and glaciers, filling of arroyos, and anchoring of dunes (pp. 119, 120).

Hansen consequently reads in his pollen profiles the same main climatic trends, the same general climatic history as has been deduced from various geological evidence in the western United States. Although no subordinate climatic variations seem to be recorded by the pollen profiles of the Northwest, there were surely in most regions additional fluctuations of several orders, that is, of different duration, amplitude, and geographic extent. Paul Sears and E. S. Deevey find that, for instance, in the northeastern United States the progressive dryness after the departure of the ice was broken, and a moist age intervened next before the age of maximum warmth and dryness (pp. 110–12). In southeastern Arizona the age of general deposition which began about four thousand years ago was temporarily interrupted by drought-induced erosion about the time of Christ and about A.D. 1300. Natural erosion would perhaps have set in again during the dry 1930's, had not overgrazing started artificial erosion in 1883. As knowledge increases, more climatic fluctuations and regional differences will be recognized, the latter mainly in precipitation and moisture.

It was to enable long-distance correlations in

[1] "Die Bedeutung der Depression der Schneegrenze für eiszeitliche Probleme," *Zeitschr. f. Gletscherkunde*, Vol. XVI (1928), pp. 70–93; see p. 75.

spite of such subordinate regional differences that Lennart von Post and I independently singled out the predominant and most widespread climatic change for time division of the Postglacial, namely, the rise, culmination, and fall of the temperature (pp. 109–11). In making correlations, it must be realized that peripheral regions were released from the ice sheets tens of thousands of years before central ones and that all the regions have had remarkably similar climatic histories despite the unequal lengths of time involved. Most correlations have to be made on a differentially sliding time scale (pp. 111, 112).

Hansen's monograph is the most comprehensive study of pollen profiles and related problems carried out in North America; it is ably done and beautifully illustrated.

ERNST ANTEVS

Charles Darwin and the Voyage of the Beagle. Edited with an Introduction by NORA BARLOW. Philosophical Library, 1946. Pp. 279; frontispiece; pls. 15; and map. $3.75.

This volume, compiled by a granddaughter, begins with a brief biography of Darwin's life to the age of twenty-one when, in 1831, he embarked on H.M.S. "Beagle," a sailing vessel of two hundred and forty-two tons and ten guns, as naturalist. At that time he appears to have been somewhat neurotic, totally lacking in ambition and completely dominated by the overpowering personality of his well-to-do father, who gave reluctant permission for the son to go adventuring only upon the intercession of an uncle, who argued that "the pursuit of Natural History, though certainly not professional, is very suitable to a clergyman."

Darwin's formal education seems to have been completely wasted—the classics and ancient history, medicine at the University of Edinburgh, and preparation for the clergy at Cambridge. Much of his spare time, however, had been devoted to the hobby of "Natural History" which was his only serious interest, but even this was not permitted to interfere with such pleasures as partridge shooting on the estate of relatives. His appointment (he received no salary and was obliged to pay his own expenses) was secured through the intercession of friends and was also influenced by the prejudice of the vessel's captain who refused to "take anyone, however good a naturalist, who was not recommended to him, likewise, as a gentleman." Altogether, young Darwin appears to have been a not too promising choice, and he certainly was singularly unprepared for such an undertaking.

The main portion of the book is divided into two nearly equal parts. The first consists of thirty-eight letters mostly written to his sisters in England and mainly unpublished previously. The letters are largely devoted to personal matters and briefly describe some of his impressions and activities principally in South America. They are of most interest because here and there they give hints of his gradual maturity, increasing self-reliance, and the steady growth of his interest in natural history and especially geology. Obviously, he returned home after five years a different man with a serious purpose in life that previously had been lacking.

The second part consists of a selection of more or less indiscriminate extracts from twenty-four field notebooks in which Darwin jotted down his observations on all subjects. These notebooks have not been published, but they furnished the basis for his *Journal*, which appeared in several editions, and the three-volume account of his geological studies. The extracts from these notes are connected by comments and interpretations by the compiler. Many of them are so abbreviated as to be almost unintelligable to the reader, and the discontinuity of the subject matter is so great as to suggest that Darwin's observations were all made in the most hit-or-miss fashion. The Introduction states that the notes are largely geological and that from half to nearly nine-tenths of the entries in different books are devoted to this subject. Perhaps because the compiler had little or no knowledge of geology, these parts have been slighted in the selection, and thus what may have been his most significant observations have been omitted.

This book is neither a very interesting nor important contribution to Darwiniana, and similar writings of a lesser man could never find a publisher. It contains little that is not better presented in Darwin's own volumes or in the previously published Darwin letters. Illustrations consist of fifteen plates of rather poorly reproduced drawings, a portrait of Darwin and one of his sisters as children, and a map showing the route travelled by the "Beagle."

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STRUCTURAL PETROLOGY OF THE BARABOO QUARTZITE

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ABSTRACT

The microstructures of the Baraboo quartzite of Columbia and Sauk counties, Wisconsin, have been studied from oriented hand specimens, polished sections, and thin sections. Fabric axes, determined from the pole patterns of the deformation lamellae, vary consistently over the region and are not related to the bedding or the macrofoliation. Because of this variation, the deformation lamellae have been assigned to a deformation later than the major folding of the region. In addition, the character of the deformation lamellae themselves and the type of pole pattern have been used as a qualitative index of the intensity of the deformation. Microfractures are common features of the quartzite; they exhibit a strong preferred orientation and are geometrically related to the deformation lamellae. Quartz deformation bands are described, named, and correlated with bedding-plane shear. Quartz optic-axes fabrics are weak and difficult to interpret.

INTRODUCTION

The present work is a field study of the outcrops and laboratory examination of oriented hand samples, polished sections, and thin sections of the Baraboo quartzite of Columbia and Sauk counties, Wisconsin. It is not my purpose to attempt a solution of the fundamental problem of the dynamics of quartz deformation or the mechanics of tectonic movements but rather to shed further light on the qualitative description of the processes of deformation and their correlation with the macrostructures observed in the field.

GENERAL GEOLOGIC DESCRIPTION OF THE BARABOO REGION

The Baraboo quartzite crops out in Columbia and Sauk counties, Wisconsin. Its structure (Fig. 1) is a complex canoe-shaped synclinorium, whose major axis strikes approximately N. 70° E. and is roughly 25 miles east-west and 9 miles north-south. The outcrop area can be conveniently divided into a north, a south, and a west range. In the north range the bedding is mainly vertical, striking east-west, with the tops of the beds to the south; in the south range it dips 20°–40° to the north, striking N. 70° E.; the much shorter west range is an area of complex, rather right, east-west folds, plunging eastward. In the western half of the south range, a series of gentle folds, trending N. 60° E., truncates the larger structure.

The Baraboo area is of considerable geologic interest because of excellent exposures of axial-plane cleavage, drag folds, etc. Numerous papers concerning the various aspects of the region have appeared in the literature. The most comprehensive treatise on the area and one with considerable emphasis on the structure of the quartzite ranges ap-

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Imperial Agricultural Research Institute

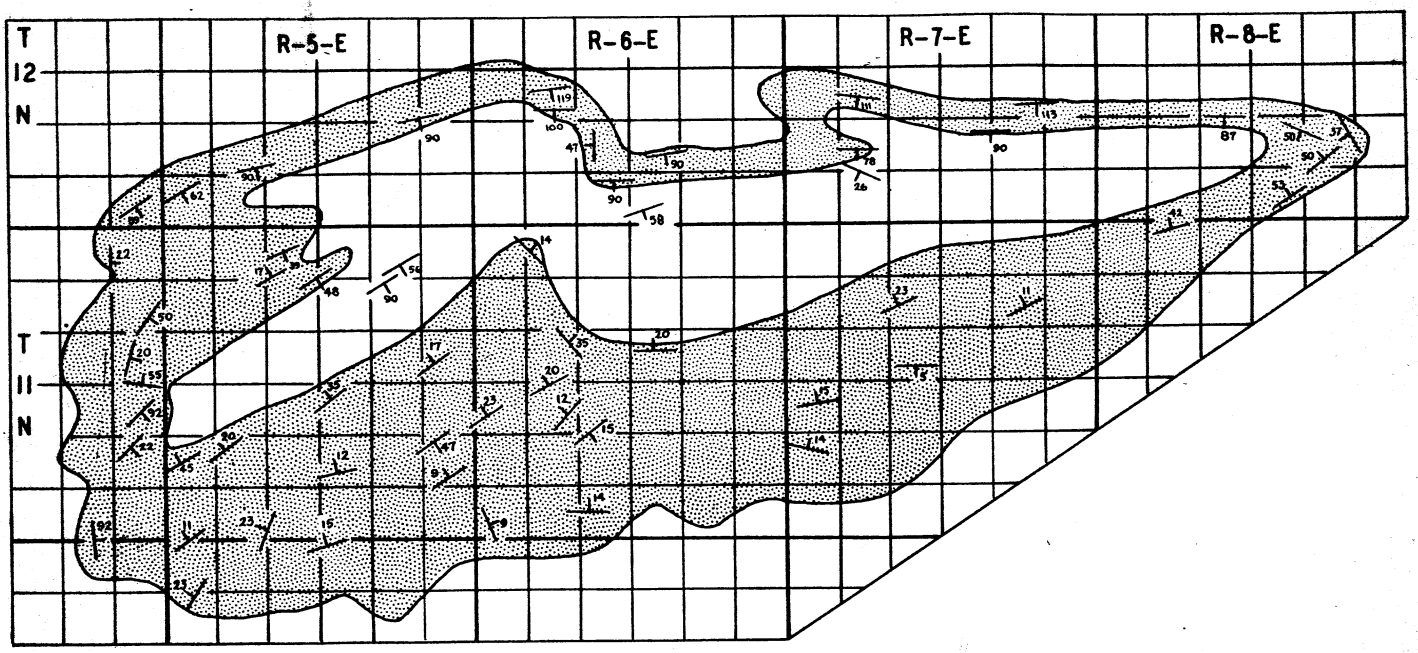

FIG. 1.—Generalized outcrop area of the Baraboo quartzite (modified from A. Leith), showing bedding orientation for selected representative locations. The four outcrops in the interior of the basin are possible outcrops of the Dake quartzite.

peared in 1904 by Samuel Weidman.[1] For the details of the region reference must be made to Weidman's work. It is sufficient to say that Weidman correctly placed the Baraboo quartzite in the Huronian, described the structure as a complex synclinorium, and divided the Huronian sediments into three formations: the Baraboo quartzite, the Seeley slate, and the Freedom iron formation. The latter two formations, which are above the Baraboo quartzite, do not crop out in the region and were described by Weidman from mine workings.

Andrew Leith[2] restudied the earlier unpublished commercial reports of C. L. Dake and in 1935, on the basis of Dake's descriptions of drill cores, divided the Baraboo Huronian sediments into an upper and lower part. Leith's pre-Cambrian column is as follows:

Algonkian	Upper Baraboo series	Rowley Creek slate Dake quartzite
	Unconformity (?)	
	Lower Baraboo series	Freedom formation Seeley slate Baraboo quartzite
	Unconformity	
Archean		Rhyolite, tuff, granite (some granite may be intrusive into the Baraboo series)

A short description of the formations of the lower Baraboo series, taken from Weidman's[3] publication, is given below:

The Baraboo quartzite exceeds 4,000 feet in thickness and is composed almost entirely of monocrystalline sedimentary quartz grains with varying admixtures of fine-grained sericitic material. Interstitial iron oxide imparts a reddish color to much of the formation. Detrital hematite and zircon grains are locally important. All large quarries in the Baraboo quartzite exhibit thin layers of almost pure sericitic material, varying in thickness from a few centimeters to several feet, interbedded with massive quartzite. Some of these thin beds are pinched into lens-shaped bodies. A poorly developed basal conglomerate is present in one known locality. Intraformational pebble bands of vein quartz are common features of the massive quartzite.

The Seeley slate is not known to crop out in the region and is recognized only in mine workings and drill holes. Its material is observable at present on the dumps of old mines and is a finely banded, soft slate, exhibiting excellent axial-plane cleavage. The thickness of the formation is unknown.

The Freedom formation is the iron-bearing formation of the region. It does not crop out in the area. Weidman describes the formation as composed of slate, chert, dolomite, iron ore, and all

[1] "The Baraboo Iron Bearing District," *Wis. Geol. Nat. Hist. Surv., Bull. XIII, Econ. Ser. No. 8* (1904), pp. 1–190.

[2] "The pre-Cambrian of the Lake Superior Region, the Baraboo District, and Other Isolated Areas in the Upper Mississippi Valley," *Ninth Ann. Field Conf. Kan. Geol. Soc.* (Wichita, Kan.: Kansas Geological Society, 1935), pp. 320–32.

[3] Pp. 22–29 of ftn. 1.

gradations among these types. The thickness of the formation is unknown but is estimated by Weidman to exceed 1,000 feet.

Leith,[4] utilizing Dake's data, named and described the formations of the Upper Baraboo series as follows:

> Rowley Creek slate: This slate does not outcrop. . . . . It occupies the middle of the synclinal trough. . . . . Its thickness varies between 40 and 149 feet. . . . .
>
> Dake Quartzite: This formation was encountered at 42 diamond drill holes. . . . . It varies in thickness up to a maximum of 214 feet. . . . . Lithologically, it consists of a coarse grained quartzite which contains a large amount of sericite and chlorite as a matrix to the quartz grains. A large portion of the formation is coarsely conglomeratic with large, and sometimes angular, pebbles. . . . .
>
> The quartzite is not known to outcrop, but the writer believes that one, and possibly several, surface exposures formerly thought to be Baraboo quartzite are of Dake quartzite. . . . .

Leith points out that two features of the structure indicate an unconformity between the Upper and the Lower Baraboo series. The drill records show that the Dake quartzite bevels the underlying Freedom formation; and, if the surface outcrops of the Dake are correctly identified, they dip at gentler angles than does the underlying exposed Baraboo quartzite. An example of the latter is Sample 55, Figure 14, which is the "type" surface exposure of the Dake. The bedding here dips approximately 25° southward, while a short distance to the north the bedding in the Baraboo quartzite is roughly vertical (the symbols used in all illustrations are explained in Fig. 2).

An attempt was made in the present study to determine fabric differences between the Dake and the Baraboo quartzites in the above outcrops in the hope that fabric differences would show the unconformity. No evidence was obtained from the mircofabric studied either to substantiate or to refute the presence of the unconformity. The fabric axes determined from the pattern of the poles of the deformation lamellae (discussed in detail later in the paper) of both outcrops are similar.

Another problem associated with the structure is the determination of the number of deformations that the Lower Baraboo series has undergone. To begin with, it appears unlikely that the complexities of structure encountered in the Baraboo quartzite could be derived in one simple deformation. Evidence is submitted that the Lower Baraboo series has suffered at least two deformations. If the angular unconformity between the Upper and the Lower series is admitted, then the region has been subjected to at least three deformations since Lower Baraboo times.

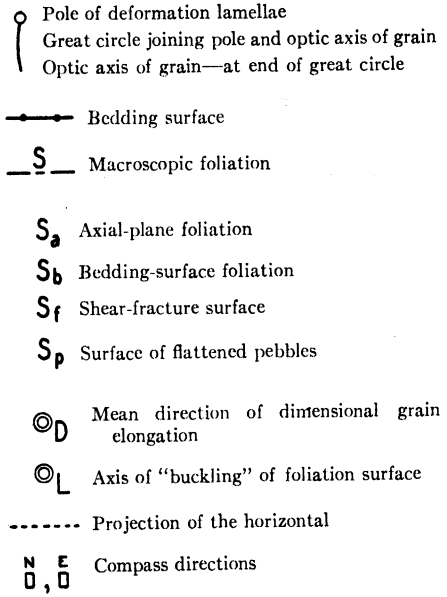

FIG. 2.—Symbols used on all maps and projections

[4] Pp. 329–30 of ftn. 2.

Volcanics, basal to the Baraboo series, occur locally around the perimeter of the synclinorium. These outcrops were not studied in detail and will not be discussed in this paper.

Resting unconformably on the Baraboo series are Upper Cambrian sands. They are roughly horizontal and crop out both in the interior of the synclinorium formed by the Baraboo series and on all sides. Several small-scale faults are found in these Cambrian sands in the interior of the synclinorium. The displacement along the fault planes can be measured in inches, the strike is roughly east-west, and the downthrow side is toward the interior of the basin. For these reasons the Paleozoic faulting is attributed to differential compaction of the sediments, and no major Paleozoic or later deformation is recognized.

The difficulties in determining the field structural relations in the Baraboo quartzite are the absence of any traceable horizon within the quartzite, the difficulty in recognizing tops and bottoms of bedding surfaces, and, in many examples, difficulties even in determining bedding surfaces. The quartzite varies tremendously in its degree of deformation over small distances. Many massive outcrops exhibit apparently undeformed ripple-marked surfaces, while a few tens of feet away stratigraphically, in more sericitic layers, large-scale drag folding and pronounced boudinage occur.

An enumeration of the principal macrostructural features follows:

1. Bedding surfaces are identified both where the change in lithology with respect to amounts of admixed sericite is gradational and where it is abrupt. Cross-bedding surfaces are easily confused with true bedding in small outcrops, and small errors in the orientation of the bedding surfaces are expected.

2. Axial-plane foliation is pronounced in the thicker layers, which contain considerable amounts of sericite. It is found only in the north range, where the bedding is close to vertical. The orientation of this foliation is roughly N. 60°–80° E., dipping 40°–60° N.

3. Bedding-plane foliation is found in the more massive layers in the north range and is the principal foliation of the south range.

4. Shear-fracture surfaces can be divided into the following types:

a) Shear fractures in massive quartzite that are observed to pass into axial-plane foliation in the more sericitic layers.

b) Chevron folding of the axial-plane foliation is locally observed. This type of structure has been described by Barrow,[5] who calls these acute plications "buckling." Fracture at the crests and troughs of the buckling and small displacements along the fracture surfaces are readily observable. The amplitude and wave length of the buckling observed varies between 0.05 and 2.5 cm.

c) Shear fractures, discontinuous and continuous, in the more massive members transcend all other structures in the south and west ranges. These are roughly vertical and trend east-west. The formation of these shear fractures was apparently subsequent to all major deformation.

5. Metamorphic surfaces consisting of flattened pebbles are prominent in the outcrop at the locality of Sample 55, Figure 14. This is a quartzite outcrop which Leith said might be an exposure of the Dake formation.

6. Macrolineation is locally evidenced by the following factors:

[5] "On Buckled Folding," *Rept. Brit. Assoc. Adv. Sci.*, Vol. LXXXII (1912), pp. 473–74.

a) Intersection of axial-plane foliation and bedding
b) Intersection of shear fractures and bedding
c) Crests and troughs of the buckling
d) Slickensides

7. Quartz veining of several forms is recognized:

a) In the north and west ranges, veins of quartz varying in thickness from a few millimeters to about 60 cm. are common. These veins follow joint patterns and transect all metamorphic structures.

b) In highly deformed quartzites, deformed quartz veins are present. Some of these deformed veins are cut by planar veins.

c) Ladder veins of quartz are found in the competent members of large-scale drag folds. They can be interpreted as vein filling of tension fractures which are roughly perpendicular to the bedding.

d) S-shaped gash veins of quartz are very common. They have not been interpreted and have not been related to the regional structure

8. Large-scale drag folds are found in three outcrops near the top of the exposed Baraboo formation in the south range, near where Sample 20, Figure 14, was collected. Boudinage is well marked, both parallel and perpendicular to the dip of the bedding.

9. No large-scale faults are recognized in the field. Leith[6] places a north-south fault between Samples 58 and 38 of Figure 14. This fault was apparently based on drill records. If a large-scale fault is present, it is not represented in the pattern of the deformation lamellae, because the fabric axes determined from the pattern of these lamellae are similarly oriented in both the above samples, as well as in those samples farther west from the supposed fault. (Fig. 14.)

[6] Pp. 329 of ftn. 2.

Consistently throughout the area, roughly vertical, north-south shear joints or perhaps shear fractures are encountered. Slickensides on these surfaces are rare but are found frequently enough to assume some north-south adjustments by fracturing. It is interesting to note that these surfaces are subparallel to the *a*–*b*-fabric planes determined from the deformation lamellae.

MICROSTRUCTURES

I. SINGLE GRAINS

The principal intragranular microstructures include the following:

1. Secondary quartz enlargement of the original quartz grains is visible in many thin sections. The original boundary is visible because a thin film of impurities at the original grain surface is incased by the secondary enlargement. The above is true only in samples containing small amounts of sericite. Samples containing large amounts of sericite, however, rarely exhibit original boundaries of the quartz grains, but each grain is completely surrounded by sericite grains. These sericite grains appear to penetrate the boundaries of the quartz grains as if sericite had replaced quartz. An original boundary is rarely observed, but when it is, it can be seen that the original quartz grain has grown by secondary quartz enlargement and has incased the adjacent sericite grains.

2. Deformation, or Böhm, lamellae are frequently observed structures within individual grains. They commonly occur in the central portions of the grains and appear as a set of closely spaced subparallel planes which are slightly lens-shaped in cross section. Infrequently, two sets of deformation lamellae are found in the same grain. A historical summary of previous work on deformation lamellae and

a careful description of them has been given by Earl Ingerson and O. F. Tuttle.[7]

3. Undulatory extinction is present in nearly all quartz grains observed. The magnitude of this change in optic orientation varies tremendously from grain to grain in one sample and in different samples. In addition to this continuously varying position of the optic axis in a particular grain, discontinuities occur which divide the grain into bands subparallel to the optic axis. Anna Hietanen[8] has extensively described deformed quartz grains from the Finnish quartzites, and she considers the undulatory extinction as being due to bend-gliding along the basal pinacoid and the discontinuities subparallel to the optic axis as actual rupture planes.

4. Undulatory extinction discontinuities with apparently random orientation are observed as thin bands across the grain. They have a slightly different optical orientation and/or index of refraction from the rest of the grain and are, therefore, seen only when the grain is very close to an extinction position. Two sets of bands, whose traces tend to be perpendicular, are present in most grains showing the banding. These bands agree in appearance with structures in vein quartz observed by Sidney Adams[9] and are termed "deformation bands," following the usage in metallurgy.[10]

5. Sutured boundaries between quartz grains are common. An examination of variously deformed quartzites shows a wide variation in the character of the sutured boundaries. The individual sutures are definitely different in different samples; they are not necessarily perpendicular to the interface between grains; they vary in an individual grain according to the optic orientation of the neighboring grain, which makes the interface; and they exhibit marked differences in the amount of interpenetration. In examples in which the interpenetration is large, the sutured parts of both grains differ considerably in optical orientation from their parent-grains. All variations between no change in optic orientation of the sutured parts of the grains to considerable change are observed. This suturing of grain boundaries and then rotation of the sutured parts with recrystallization could be major processes in the breakdown of large grains. In addition to the above, differences in the amount of interpenetration of the sutures are observed in variously oriented sections of the same sample. This observation indicates that the character of the sutures is probably related to the major stress pattern. No investigation of suturing other than the above general observations has been made in this study.

Much less frequent than suturing is the "pressing-in" of one side of a particular grain into an adjacent one. In some examples it is possible to observe the complete original boundary of one of the grains and the partial original boundary of the other grain.

6. Breakdown of the original grains by deformation can be observed as having taken place

a) Along planes subparallel to the optic axis, where the original grains break down into needle-like fragments exhibiting strong undulatory extinction.

[7] "Relations of Lamellae and Crystallography of Quartz and Fabric Directions in Some Deformed Rocks," *Amer. Geophys. Union Trans.*, Vol. XXVI (1945), pp. 95–105.

[8] "On the Petrology of the Finnish Quartzites," *Bull. Comm. géol. Finlande, No. 122* (1938), pp. 1–118.

[9] "A Microscopic Study of Vein Quartz," *Econ. Geol.*, Vol. XV (1920), p. 648.

[10] Charles S. Barrett, *Structure of Metals* (New York: McGraw-Hill Book Co., 1943), pp. 305–7.

b) Along the sets of deformation bands, where the whole grain breaks down into small angular units that are roughly equant in cross section and vary slightly from one another in optic orientation. The end-result of this breakdown could result in the texture called "crushed quartz" by R. E. Fellows.[11]

c) At the boundaries among grains, where small sutured fragments are detached from their parent-grains.

d) In the interior and at the boundaries among grains, where recrystallization like that illustrated by J. A. Hedvall[12] occurs, that is, growth of new crystals at the expense of the old ones.

7. Small subhedral grains are found in the interstices among the larger grains. They show no trace of a rounded surface, have very irregular but equant shapes, and appear as if they had grown around new nuclei, that is, had recrystallized.

8. In some of the quartzites that contain negligible amounts of sericite and have suffered extensive deformation, as shown by the partial breakdown of the larger grains, very elongate quartz grains are observed. Rarely, one of these has an original sedimentary boundary, indicating that it formed from an original grain and was not the result of growth from a new nucleus or of coalescence of several grains. An elongate grain commonly exhibits marked discontinuities in the undulatory extinction subparallel to the optic axis when it is subparallel to the dimensional elongation. On the other hand, if the grain is dimensionally elongated subperpendicular to the optic axis, it may show well-developed deformation lamellae in addition to the discontinuities in the undulatory extinction. Samples of quartzites containing large amounts of sericite display little elongation of the original quartz grains of the sand size. These grains have reacted to the stress like a rigid body in a fluid medium.

[11] "Recrystallization and Flowage in Appalachian Quartzite," *Bull. Geol. Soc. Amer.*, Vol. LIV (1943), p. 1414.

[12] *Reaktionsfähigkeit fester Stoffe* (Leipzig: Verlag von Johann Ambrosius Barth, 1938), p. 53.

II. POLYCRYSTALLINE AGGREGATES

The principal intergranular microstructures of the Baraboo quartzite include the following:

1. Planar structures passing through more than one grain without deviation at the grain boundaries are commonly observed. Some of these are marked by thin veinlets of very fine-grained sericite. In most examples, however, they are discontinuous planes of inclusions. In all examples, continuous or discontinuous, these planes show a pronounced parallelism. More than one set of such structures can be identified in some samples. These planar structures are termed "microfractures."

2. Quartz microveins are common. They occur as planar structures and might be thought of as having grown along the microfracture surfaces described above. In addition, quartz microveins are found that could be described as "contorted." Both types have been found in the same thin section, with the planar veins cutting the contorted veins.

3. Microfoliation surfaces are visible in samples containing large amounts of sericite as concentrates of sericite cutting across relic bedding but not cutting through original quartz-grain boundaries.

4. Original sedimentary structures commonly observed in thin sections are the following:

a) Laminae of heavy mineral concentrates are recognizable in a large number of samples. These laminae are valuable

field aids for the determination of the sedimentation surface. They rarely exceed a few millimeters in thickness and are composed exclusively of hematite and zircon grains. Sufficient zircon grains could be found for a regional study of the zircon fabric.

b) Graded bedding is prominent on a microscale, and in most examples one graded unit represents one cross-bedding lamina.

c) Ripples and truncated ripples are visible in thin section by slight changes in grain size or by variation in abundance of hematite, zircon, or sericite.

d) Some polycrystalline (sedimentary) quartz grains are present in the more conglomeratic phases of the Baraboo quartzite. Some of the individual grains show a much higher degree of metamorphism as well as a greater variation in degree of metamorphism from grain to grain than is true for monocrystalline grains of the same size in the same sample. This variation indicates a source region or regions in which the quartzites differed widely in degree of metamorphism. As will be shown later in this paper, the region may be as small as the outcrop area of the Baraboo quartzite and still show this wide range in degree of metamorphism.

DETAILS OF THE MICROSTRUCTURE

I. DEFORMATION LAMELLAE

The deformation lamellae observed in the Baraboo quartzite agree in the main with those described by Ingerson and Tuttle.[13] Contrary to their findings in the Ajibik quartzite and in a biotite gneiss from near Washington, D.C., however, deformation lamellae are common in the Baraboo quartzite and are relatively easy to observe in the majority of examples. In samples containing so much sericitic material that the individual quartz grains are not in contact with one another, however, the readjustments during deformation have been made by the sericite grains leaving the quartz apparently undeformed and without deformation lamellae.

Various interpretations have been given to the deformation lamellae, but, following Ingerson and Tuttle, they can all be classified as: (1) planes of liquid inclusions, (2) healed fractures, and (3) translation glide-planes. Their exact nature is not known; their relation to quartz structure is obscure; and their orientation with respect to the major-stress axes is based on field evidence. Deformation lamellae have not been observed in laboratory deformation of quartz.

H. W. Fairbairn[14] measured the angle between the poles of the deformation lamellae and the optic axes of the same grain and found a total variation of from 3° to 85° with maxima at 18° and 27°. He interprets the deformation lamellae as translation glide-planes containing the glide-line [*m*:*r*].

Ingerson and Tuttle[15] found the angle between the optic axis and the pole of the deformation lamellae to vary with the angular distance from the optic axis of the grain to the *B*-fabric axis ("*B*" is written "*b*" in this paper). In addition, they tabulated the solid angle formed by the planes containing the optic axis and poles to two sets of deformation lamellae from the same grain. If deformation lamellae are translation glide-planes containing the glide-direction [*m*:*r*], then this angle is 60°. No evidence of this 60° angle was found.

[13] Pp. 95–96 of ftn. 7.

[14] "Quartz Deformation and Crystal Structure," *Amer. Min.*, Vol. XXIV (1939), pp. 351–68.

[15] P. 101 of ftn. 7.

The deformation lamellae do form more frequently at low angles to the basal pinacoid. The distribution of angles (optic axis to pole of deformation lamellae) of Ingerson and Tuttle and of Fairbairn, tabulated in Figures 10 and 11 of Ingerson and Tuttle,[16] were substantiated by measurements on the Baraboo quartzite. In addition, the perpendicular distance between individual lamellae was measured. The spacing varied between 2.5×10^{-4} and 17×10^{-4} cm., with a mean value of 8.3×10^{-4} cm. These values agree in magnitude with the observed spacing of slip-planes in metallic crystals.

Ingerson and Tuttle relate the deformation lamellae to the fabric axes; that is, the intermediate-strain axis is parallel to the lamellae (or the lamellae poles lie in the *a*–*c*-fabric plane). Ingerson and Tuttle have further pointed out that if the lamella pole and the optic axis of quartz are plotted for an individual grain and are joined by a great circle, the pole to the lamella always lies between the optic axis and the *a*-fabric axis (Fig. 3), that is, the great circles joining optic axes and deformation-lamellae poles always "point" to the *a*-fabric axis.

[16] P. 99 of ftn. 7.

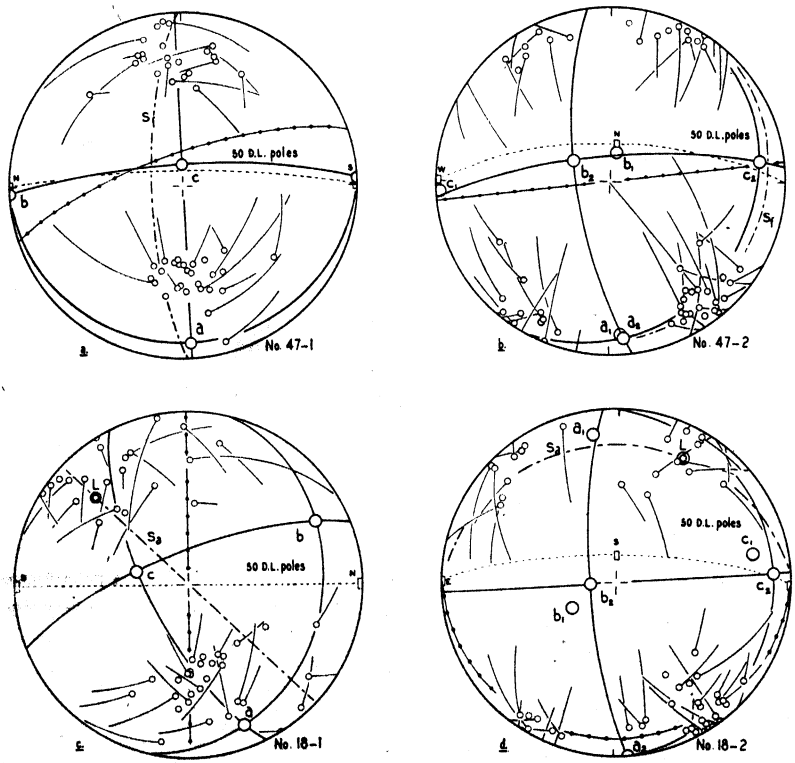

FIG. 3.—From Sample 47 two perpendicular sections have been cut, 47-1 and 47-2. The *a*-, *b*-, and *c*-fabric axes were determined independently in each section from the pattern of the poles of the deformation lamellae. The fabric axes determined from 47-1 are plotted on the projection for 47-2 and labeled as a_1, b_1, c_1. The above also applies to the two sections from Sample 18. The top of the bedding is to the south in Sample 18.

No consistent relationship was found in the Baraboo quartzite between the *a–b*-fabric plane determined in the field from the macrofoliation surfaces and the fabric axes determined from the pattern of the deformation lamellae. To avoid confusion of terminology, the *a*-, *b*-, and *c*-fabric axes are directions determined according to Ingerson and Tuttle from the pattern of the deformation lamellae. Megascopic foliation surfaces in all examples are here designed as *S*-surfaces, with subscripts to denote type of *S*-surface. Figure 14 shows the variation in the fabric axes as determined from the deformation lamellae over the outcrop area of the Baraboo quartzite. The interpretation of this pattern will be discussed later.

In most samples the poles of the deformation lamellae are concentrated in two small maxima, and the *a*-, *b*-, and *c*-fabric axes can be critically located with reproducible results from variously oriented thin sections of the same sample and from different areas of the same thin sections. Two examples of this for a pair of mutually perpendicular thin sections from the same sample are given in Figure 3.

As deformation of the aggregate increases, an individual grain will be subjected to a changing-stress field. Deformation of a particular grain and deformation of its neighboring grains will alter the local stress pattern. A wide range in character, strength, and deformation of the deformation lamellae themselves is observed in an increasing-deformation field. Increasing deformation, of course, applies to a sequence of changes as observed in a series of rock specimens. In addition to the location of the *a*-, *b*-, and *c*-fabric axes, therefore, considerable evidence concerning the deformation can be obtained from the variation and character of the deformation lamellae. Ingerson and Tuttle[17] indicated the above but did not elaborate.

The following detailed observations serve to illustrate the variations in the character of the deformation lamellae and in the pole pattern of the deformation lamellae. They are from samples containing negligible amounts of sericitic material, unless otherwise indicated.

1. The samples which exhibit the most precisely oriented pattern of poles of the deformation lamellae are those whose quartz grains show slight undulatory extinction, no noticeable dimensional elongation, distinct original boundaries with secondary enlargement, and rare other intragranular discontinuities. From the above, it appears that in an increasing-deformation field these lamellae are the first deformation structures to appear. Many examples can be found in which the deformation lamellae cross the boundary from the original grain into the region of secondary enlargement. Most of the deformation lamellae, therefore, followed the secondary enlargement. The exact relations between the development of undulatory extinction and the formation of deformation lamellae is difficult to evaluate, but grains exhibiting deformation lamellae also show some slight undulatory extinction. Sample 47, whose pole pattern of deformation lamellae is shown in Figure 3, is an excellent example of the above.

2. Curved deformation lamellae are found in samples exhibiting strong undulatory extinction, which has discontinuities subparallel to the optic axis, seldom observed original grain boundaries, and a noticeable grain dimensional elongation. The orientations of the great circles joining the optic axes and the poles of the deformation lamellae vary from the

[17] P. 105 of ftn. 7.

ideal examples shown in Figure 3. Figure 4, Sample 9, is the pole pattern of a sample showing the above features.

3. One large diffuse maximum of the poles of the deformation lamellae and several smaller minor maxima are shown by samples exhibiting the partial breakdown of the original grains to smaller elements, rare original grain boundaries, and a pronounced grain dimensional elongation. The *a*-, *b*-, and *c*-fabric axes are difficult to locate from the pole pattern. The deformation lamellae in the elongate grains are oriented about 30° to the elongation (Fig. 5, Sample 28). The unequal development of the pole maxima of the deformation lamellae could be explained in the following ways:

a) The deformation lamellae tend to form subperpendicular to the *c*-axis of the grain. With certain patterns of preferred orientation of the *c*-axes, one set of deformation lamellae would be more common than the other. Diagrams showing a definite *c*-axis preferred orientation can possibly be correlated with the deformation lamellae pole pattern (Fig. 5).

b) The deformation lamellae are re-

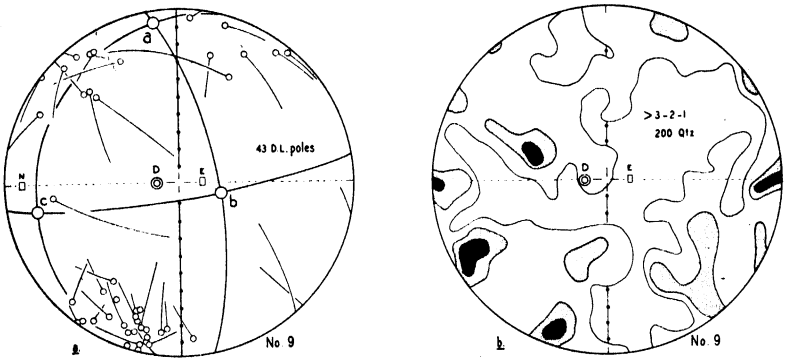

FIG. 4.—Pattern of poles of deformation lamellae in which one maximum is stronger than the other. A weak optic-axes fabric from random grains indicates a possible correlation.

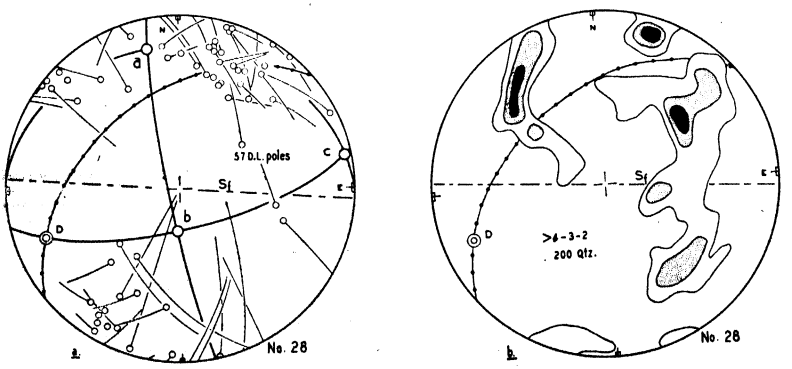

FIG. 5.—One large diffuse maximum of poles of the deformation lamellae, one or possibly two small maxima, and many aberrant lamellae make the choice of the fabric axes difficult. The pronounced dimensional elongation and the definite preferred orientation of the optic axes is not directly correlative with the deformation-lamellae pattern.

lated to the macroshear surfaces, one of which is better developed than the other. No constant relation was found between macrofoliation surfaces and the planes of deformation lamellae.

c) The deformation lamellae are related to the dimensional grain elongation, i.e., metamorphic dimensional grain elongation, which, in turn, is related to the optic-axes fabric. In rocks showing the most highly oriented and symmetrical pattern of poles of deformation lamellae, the deformation lamellae appear to form subperpendicular to the long-dimensional axis of the grains. This is "according to Hoyle" because grains which are nonequant suffer, on the average, the greatest bending parallel to their long dimension. If actual movement, either by gliding or by fracture, takes place along the deformation lamellae, then, under continued deformation, the individual grains are elongated in a direction in the plane of the deformation lamellae. With infinite elongation the greatest dimensional elongation of the grains would become parallel to the set of deformation lamellae causing the elongation. This hypothesis is crudely verified by measurements of the angle between the elongation and the deformation lamellae in specimens exhibiting various amounts of grain elongation. In addition, the grain elongation can be expressed as a flattening, which can be taken as two principal directions of elongation. The better-developed system of deformation lamellae would then logically be subparallel to the greater axis of elongation. In only one example—Sample 55, Figure 6—flattening of the quartzite grains was observed, and the above relation was not verified. In this example, however, one set of deformation lamellae is perpendicular to the flattening and one subparallel.

The logical conclusions are twofold. In very elongate grains in which the deformation lamellae are oriented at small angles to the dimensional elongation, the observed dimensional elongation could have been accomplished by gliding along the deformation lamellae. On the other hand, if the deformation lamellae are roughly subperpendicular to the long-dimensional elongation of the grains, then the deformation lamellae were formed after the event which caused the grain to be dimensionally elongated.

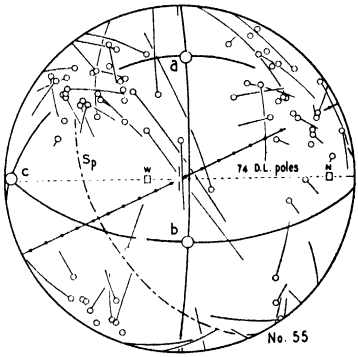

FIG. 6.—Conglomeratic quartzite (Dake?) in which the grains are flattened. In the surface of flattening, S_p, the greater elongation approximately coincides with the bedding. The deformation lamellae and the flattening are possibly related in that one pole maximum coincides with the surface of flattening and the other is approximately perpendicular.

The above interpretations are undoubtedly related but are difficult to demonstrate because neither the orientation of the forces applied nor the number of deformations is known and also because the measurement of average grain elongation in the thin section is difficult and based on the assumption that the grains are cylindrical in shape.[18] Since the major axis of metamoprhic dimensional grain elongation apparently lies

[18] Kurt E. Lowe, "A Graphical Solution for Certain Problems of Linear Structure," *Amer. Min.*, Vol. XXXI (1946), pp. 425–34.

close to the bedding surfaces (Fig. 14, Samples 9, 55, 35, 28, 3), it seems logical to assume that the stress causing the elongation has acted along the bedding surfaces. The pattern of deformation lamellae should then be related to the bedding and to the axial-plane foliation, which is, in turn, related to the bedding. No such relation is found consistently over the region. A possible explanation is that the pattern of deformation lamellae represents a later deformation imprinted

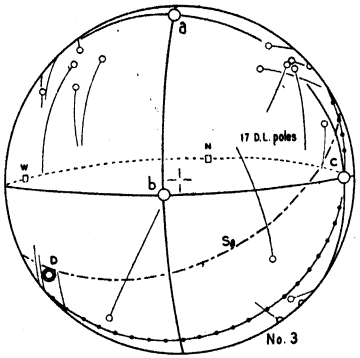

FIG. 7.—Pole pattern of deformation lamellae from a sericite schist layer. The few deformation lamellae observable adequately bring out the pattern. No relation is apparent between the pole pattern and the axial-plane foliation, S_a, observed from the outcrop, in the hand sample, and in the thin section. For another example of the same rock type, see Sample 18, Fig. 3*c* and *d*.

on an earlier pattern of metamorphic structures. Samples 3 and 18, Figures 7 and 3, are from sericite schist members of the Baraboo quartzite which exhibit marked axial-plane foliation. The fabric planes determined from the deformation lamellae do not bear any detectable relation to this foliation. In their examples Ingerson and Tuttle have demonstrated that the fabric axes determined from the deformation lamellae are related to the macrofabric axes, that is, the *a*–*b*-fabric plane determined from the deformation lamellae is parallel to the axial-plane foliation. If we accept their data, then the fabric axes determined from the deformation lamellae in the Baraboo quartzite are related to a different system of forces than those which resulted in the axial-plane foliation.

4. In several pole patterns of deformation lamellae (Samples 5 and 6 in Fig. 14; the pole pattern for Sample 6 is shown in Fig. 8), the poles form a girdle about the *a*-fabric axis. This girdle has an apical angle of about 80°. The pattern, therefore, has one degree of freedom about the *a*-fabric axis; thus the *b*- and *c*-fabric axes cannot be uniquely determined. Either of two explanations appears possible: that this pattern represents true biaxial strain or that this pattern represents two or more deformations, during which the *a*-fabric has remained essentially constant in direction but with various directions of the *b*- and *c*-fabric axes.

5. Deformation lamellae whose planes are marked by a thin film of brown material are recognized. Ingerson and Tuttle,[19] when they can detect no difference in index of refraction between the lamellae and the rest of the grain, call this type of lamella "relict." Very minute changes in the density of the brown film in one lamella can be traced through the set of the lamellae, subperpendicular to the lamellae, subparallel to the optic axis of the grain, and parallel to the discontinuities in the undulatory extinction. Hietanen describes a quartzite in which the deformation lamellae, termed "Böhm lamellae" in her paper, are curved, showing a wavelike pattern. Her diagrams[20] illustrate the change in texture of the brownish material along the lamellae, as described above, at the positions of curvature of the lamellae. Curvature of the deformation lamellae in

[19] P. 95 of ftn. 7. [20] Pp. 39–40 of ftn. 8.

the Baraboo quartzite and/or changes in the density of the brown film along the lamellae coincide with discontinuities in the undulatory extinction subparallel to the optic axis.

Hedvall and Sjöman[21] have shown that, at temperatures above the inversion point of quartz, Fe_2O_3 diffuses into the quartz lattice at a rate measurable in the laboratory. With sufficient time and perhaps in the presence of impurities, it is entirely possible that appreciable diffusion occurs at lower temperatures and that the brown films marking the deformation lamellae have originated in that manner.

6. Two sets of deformation lamellae rarely occur in a single grain. When observed, one set is more pronounced and appears to cut the other set. This observation does not necessarily imply two deformations, because two sets of deformation lamellae in the same grain could be explained as the result of local perturbations in the stress field during the same deformation.

If the relict deformation lamellae could be separated in a particular sample from deformation lamellae not showing the brown film and if the relict lamellae were cut by the nonrelict lamellae and if the pole patterns of the two types of lamellae were consistently different and if the regional configuration of the rocks could be correlated with two distinct deformation-lamellae patterns, then perhaps multiple deformations could be inferred. A lengthy attempt was made to achieve this result, but without success.

Summarizing the foregoing, it is possible to achieve a qualitative estimate of the intensity of the local deformation from the pattern of the deformation-lamellae poles and from the character of the deformation lamellae themselves. In an increasing intensity of deformation the following varieties appear:

1. *Straight deformation lamellae.*—These are usually in the central portions of the grains; and they are roughly perpendicular to the long axes of the grains; the pole pattern of the deformation lamellae is clear cut, with two well-developed maxima approximately 80° apart; the pole of a particular set of deformation lamellae always lies between the *a*-fabric axis and the optic axis of the particular grain.

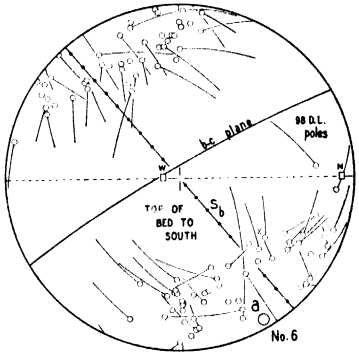

FIG. 8.—Pattern of poles of deformation lamellae in which the poles are distributed in a small circle about the *a*-fabric axis.

2. *Curved deformation lamellae.*—These lamellae appear with the curvature along planes subparallel to the optic axis and parallel to the discontinuities in undulatory extinction; the relation between the deformation lamellae and dimensional grain elongation is obscure; the pole pattern of the deformation lamellae is more diffuse than in the above examples, but the two maxima are clearly recognizable.

3. *Curved deformation lamellae as above.*—Dimensional grain elongations tend to be at small angles to the plane of the deformation lamellae; the pattern of poles to the deformation lamellae is dif-

[21] P. 212 of ftn. 12.

fuse, so that the fabric axes are determined with considerable question.

The above analysis applies only to "pure" quartzites, that is, each quartz grain is bounded by quartz grains with only small amounts of other material, such as sericite, in the interstices among the quartz grains. The stresses are then transmitted principally from quartz grain to quartz grain. As the amount of sericite, for example, is increased, part of the change in shape of the deforming aggregate is borne by the sericite. With sufficient admixture of sericite, such that the quartz grains are no longer in contact with one another, the major portion of the change in shape of the aggregate in deformation is accomplished by change in the sericite, the quartz grains acting as rigid bodies in a viscous fluid.

The study of the deformation lamellae in the Baraboo quartzite indicates the need for a better understanding of the conditions of their formation and of their exact nature. Only by producing them experimentally can the relation between the forces and the ensuing pattern be determined. If this could be accomplished, deformation lamellae should be one of the more powerful tools in the study of deformed rocks.

II. DEFORMATION BANDS

A detailed study of the variation in the orientation of the optic axis in individual quartz grains of the Finnish quartzites has been made by Hietanen.[22] She plotted the orientation of the optic axis of a particular grain on an enlarged trace of the grain outline. These plots show the angular limits of the undulatory extinction and demonstrate that the extinction varies continuously over certain limited areas, which are separated from one another by discontinuous changes in the undulatory extinction. These discontinuities, in all her examples, are subparallel to the general orientation of the optic axes of the particular grain and impart to the grain, under crossed nicols, a banded appearance. Most of Hietanen's examples are highly metamorphosed, and the grains exhibit a banded appearance over the entire grain.

In the Baraboo quartzite, however, other discontinuities in the undulatory extinction are found. It is these discontinuities, and not the ones subparallel to the optic axis, that are termed "deformation bands." Adams[23] noted similar structures in vein quartz and described them as follows:

> The latticed structure shown by inclusions may sometimes be seen under crossed nicols (Pl. XXVI., *F*). The features brought out by this photograph are: (1) each single apparent veinlet consists of a thread of quartz which disagrees in extinction position with the parent crystal; (2) the crystallographic orientation of the quartz in parallel threads, the traces of roughly parallel planes, is the same; (3) the orientation in one set of planes is not the same as in the other set.

Deformation bands are infrequent in the Baraboo quartzite, but in two examples they are common (Samples 38 and 16, Figs. 9 and 10). The shift in the optic axis from grain to band is very small, so that the bands are observed only close to the extinction position of the grain. For fifty-four measured shifts in orientation from grain to band, the mean shift was 5°.9, with a standard deviation of 4°. The mode was in the 2°–4°-class interval, however, indicating that a few large values have substantially increased the mean. In addition, there is a suggestion that the distribution is bimodal. The surprising thing about the change in orientation is that in most examples the change is usually in the same rotational sense. In

[22] Pls. I–VII of ftn. 8.

[23] P. 648 of ftn. 9.

Sample 38, Figure 9*b*, for example, the sense of rotation from grain to band indicates that the younger beds to the south have moved upward with respect to the older beds to the north. In Sample 16, Figure 10*b*, however, the relationship is not clear cut. In many examples the difficulty is in determining which part of the grain is band and which part of the grain is the supposed undeformed part.

It is not possible to determine the relation between the deformation bands and the structure of the quartz, because the spatial orientation of the bands cannot be optically determined. The traces of the bands in the plane of the thin section, however, tend to be perpendicular and parallel to the optic axes. In addition, in Sample 38, Figure 9*b*, the traces of the deformation bands are mainly at a small angle to the bedding. If the bedding surface has acted as the shearing surface, then the relation of the traces of the deformation bands to the bedding is similar to the results of experimental work of W. Riedel, which has been discussed by A. Nádai.[24]

Riedel sheared plastic clay and found

[24] *Plasticity* (New York: McGraw-Hill Book Co., 1931), p. 330.

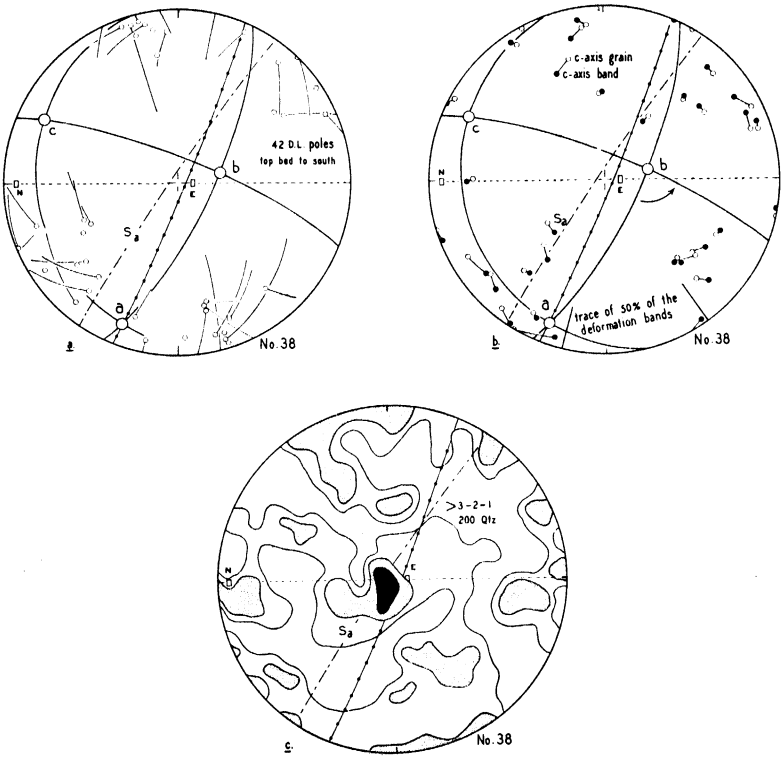

FIG. 9.—*a*, Pattern of poles of the deformation lamellae. The *a*–*b*-fabric plane is close to the axial-plane foliation determined in the field, from the hand sample, and in the thin section. *b*, Change in orientation of the optic axis from the major portion of a grain to the deformation band. All shifts are in the same rotational sense. *c*, Slight preferred orientation of the optic axes. The weak north-south girdle and the maximum at the center of the projection are reproducible.

that shearing cracks formed about 12° to the direction of shear. These are oriented similarly to the traces of the deformation bands in Sample 38, if the bedding plane is taken as the shear plane and the direction of shear is taken from the change in optic orientation from grain to band.

These bands need not be simply deformed quartz but could be a stress modification of SiO_2.

III. MICROFRACTURES

Microfractures are observed in most examples of the Baraboo quartzite. These fractures can be divided into those which are confined to individual crystals and those which pass through the aggregate of crystals without deviation at the crystal boundaries. It is the latter which exhibit a very strong preferred orientation and whose orientation pattern correlates very closely with the fabric axes determined from the pattern of the poles of the deformation lamellae. Sample 58, Figure 11, exhibits the close correlation of deformation-lamellae and microfracture patterns. The maximum of the poles of the microfractures lies but a few degrees from the *a*-fabric axis, determined independently from the deformation

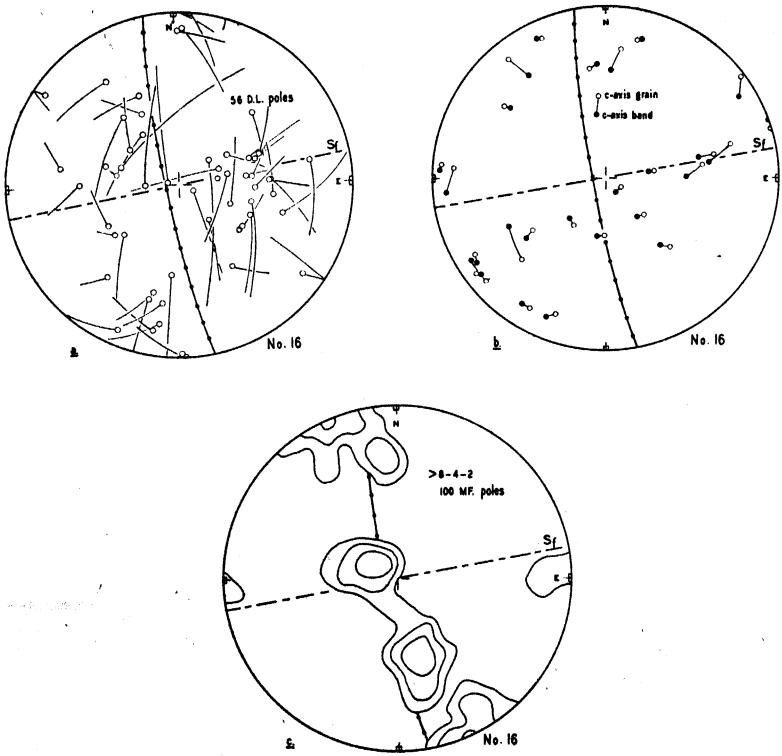

FIG. 10.—*a*, Diffuse pole pattern of deformation lamellae. The fabric axes for this sample shown in Fig. 14 are questionable. *b*, Change in orientation of the optic axis from grain to deformation band. Slight tendency for rotation to the southwest. The top of the bedding is to the east, and the younger beds have moved up with respect to the older. *c*, Poles of microfractures form a partial girdle, which is perpendicular to the shear fractures and subparallel to the bedding.

lamellae. A corresponding relation is shown in Sample 21, Figure 12. These two examples make it difficult to avoid the inference that the deformation lamellae and the microfractures are both related to the same set of external forces. In addition, since the deformation lamellae in the Baraboo quartzite are not consistently related to the bedding and the foliation, the microfractures are not so related.

The principal set of microfractures in the foregoing two examples and in all examples observed in the Baraboo quartzite are oriented very close to the *b–c*-fabric plane determined from the deformation lamellae. They are, therefore, termed, "*b-c* microfractures." Since the *a*-fabric axis is generally close to vertical, the *b–c*-fabric plane and the *b-c* microfractures are close to horizontal. In one sample, however—Sample 7, Figure 14—the *a*-fabric axis, determined from the deformation lamellae, lay close to the horizontal. The maxima of the poles of microfractures from this sample followed the above examples and coincided with the *a*-fabric axis.

In several patterns of poles of microfractures, two maxima are present. One

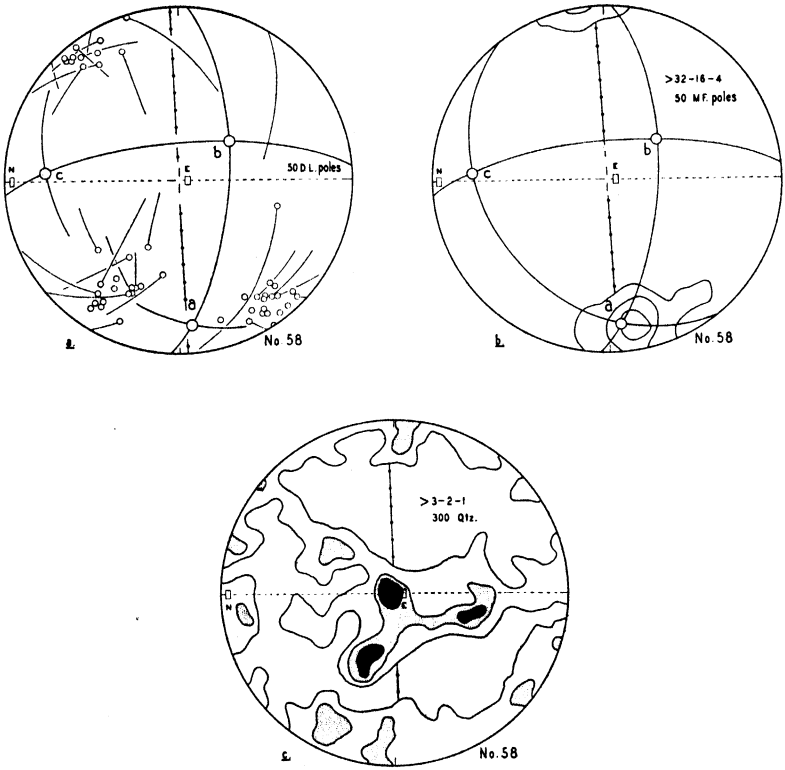

FIG. 11.—*a*, Symmetrical pole pattern of deformation lamellae, giving accurate location of the fabric axes. The sample is from a massive quartzite. *b*, The maximum of poles of microfractures is a few degrees from the *a*-fabric axis determined from the deformation lamellae. *c*, Weak north-south optic-axis girdle, partial horizontal girdle, and maxima close to axis of the weak north-south girdle.

maximum is close to the *a*-fabric axis, the other is weaker and close to the *b*-fabric axis. This type of pattern is shown in Sample 21, Figure 12, referred to above. These latter microfractures are termed "*a-c* microfractures." No examples were found in which *a-c* microfractures were present and *b-c* microfractures absent. In addition to the above two maxima, a few samples exhibit rare microfractures in the *a–b*-fabric plane.

Sample 16, Figure 10, evidences a partial girdle of poles to microfractures. O. F. Tuttle[25] in an abstract of his study noted that in the rocks of the Washington area planes of inclusions passing through grain boundaries (termed "microfractures" in this paper) are not in agreement with local structure but are uniform over 15 or 20 miles. Also, he says that, in drag folds, planes of inclusions develop in shear directions and that one set predominates. Further, he observed that poles of the inclusion planes tend to be in a partial girdle about the fold axis. The axis of the microfracture pole girdle observed in Sample 16 is approximately east-west, corresponding closely to the strike of the region and perpendicular to the ever present east-west shear fractures. It was not possible to correlate the microfractures with the deformation lamellae in this sample because of the diffuse pattern of poles to the deformation lamellae.

The brief study of the microfractures in the Baraboo quartzite indicates that microfractures can be correlated with the pattern of the poles of the deformation lamellae and that *b-c* microfractures are present if any microfractures are present. Since the pattern of the poles of the deformation lamellae do not correlate with the macroscopic foliation, the microfractures also do not so correlate.

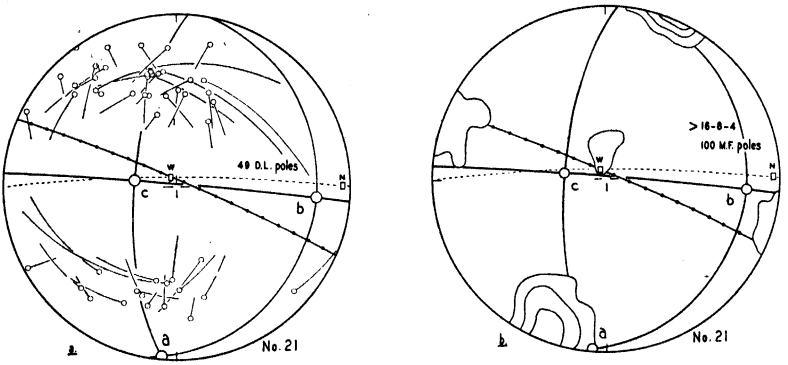

FIG. 12.—*a*, Symmetrical pole pattern of deformation lamellae. *b*, Three maxima of poles of microfractures at right angles. The principal maximum is close to the *a*-fabric axis determined from the deformation lamellae.

IV. QUARTZ AND ZIRCON OPTIC-AXES FABRIC

The results of the study of the orientation of the quartz optic axes were discouraging. Many of the patterns are not reproducible from different parts of the same thin section. Three features of the weak patterns, however, do stand out and are usually reproducible: (1) a weak north-south vertical partial girdle; (2) a

[25] "Structural Relations of Planes of Liquid Inclusions in the Crystalline Rocks of the Washington (D.C.) Area," *Amer. Geophys. Union Trans.*, Vol. XXVI (1945), p. 301.

small maximum close to the bedding and in the weak north-south girdle; (3) a small maximum close to the bedding and subperpendicular to the weak north-south girdle. In addition to the above features, it is possible in some examples to recognize a weak horizontal girdle.

The more definite examples of preferred orientation of the optic axes are found at the extreme eastern end of the outcrop area. Figure 5*b*, Sample 28, and Figure 13*a*, Sample 32, are two examples (Sample 32 is located near Sample 35, Fig. 14). No interpretation of any of the quartz-axes fabrics is attempted in this paper.

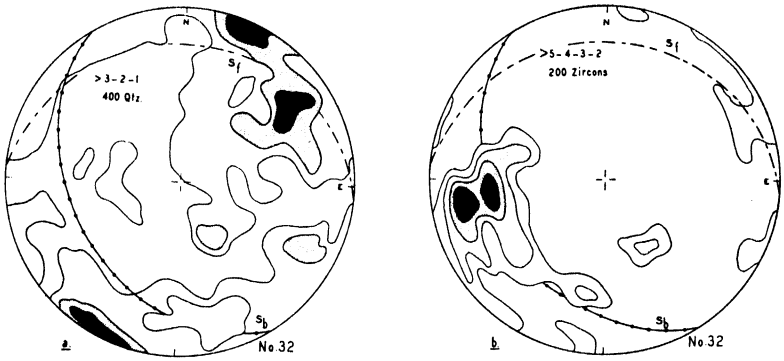

FIG. 13.—*a*, Pattern of optic axes of random quartz grains. A weak north-south girdle, dipping eastward, correlates with the pattern of Sample 28, Fig. 5*b*, which is on the south limb of the nose of the syncline. *b*, Pattern of optic axes of the zircon grains.

An examination of the outcrops, hand samples, and thin sections in the eastern half of the region shows that the deformation and the degree of preferred orientation increase from west to east toward the nose of the synclinorium. Comparisons among the samples can be made only among those which contain roughly the same admixture of sericitic material. This will be restated in the summary.

A flood of zircon grains was found in Sample 32, Figure 13*b*, located near Sample 35, Figure 14. The zircons occur in hematite laminae which are interbedded with laminae of quartz grains. The laminae are roughly 1 mm. thick. A fabric diagram was constructed for the optic axes of 200 zircons. One large maximum is located along the bedding plane. The individual zircons are fractured, but all exhibit rounded boundaries, while the quartz grains, a fraction of a millimeter away, rarely exhibit an original boundary and have been dimensionally elongated. The zircons are embedded in hematite (probably originally detrital iron oxide), and their orientation could have been accomplished by the dimensional orientation of the rigid zircons in a softer medium—i.e., hematite. This interpretation, of course, implies a correlation between the optic and the dimensional grain axes.

V. DIMENSIONAL GRAIN ELONGATION

The true spatial distribution of the axes of grain elongation is impossible to determine uniquely from thin-section studies. Lowe[26] determined a mean lineation from variously orientated outcrop surfaces of granite by assuming that the

[26] Ftn. 18.

grains are cylindrical in shape. In addition, he says that he has applied the same procedure to variously oriented thin sections of the same sample apparently by noting the modal direction in each thin section. These assumptions are definitely not true for most minerals developed through dynamic metamorphism; and, even if they were true, it is not possible to determine the actual frequency distribution from the frequency distribution of its projections. The above technique, however, does give an indication of the most frequent elongation if the grains are roughly cylindrical in shape and was applied to a few examples of the Baraboo quartzite.

Only five clear-cut examples of grain elongation were encountered in the samples studied (Fig. 14). In all except one example, the major elongation of the grains trends in an east-northeast direction. This direction is also the approximate strike of the bedding and the axial-plane foliation. Most workers in structural petrology consider that the grain elongation is usually parallel to the intermediate strain axis of the deformation, that is, to the fold axis. If that is true in the Baraboo region, the grain elongation is the result of the major deformation, with a fold axis striking roughly east-northeast.

The relation between the grain elongation and the pattern of deformation lamellae has been mentioned before in connection with the deformation lamellae. The writer does not believe it feasible to determine the exact relations between deformation lamellae and elongation from the Baraboo region, because we are not sure of the number of deformations, the orientation of the forces causing the deformation, etc. A laboratory study and/or a field study in a region in which the boundary conditions are well known should indicate the solution.

SUMMARY

1. A study of Figure 14 indicates that the strike of the *a–b*-fabric plane, determined from the pattern of the deformation lamellae, varies over the synclinorium and, with very few exceptions, cuts across the strike of the bedding and foliation. If these fabric directions are found in other regions to correlate with the foliation, as has been reported by Ingerson and Tuttle, then the patterns of deformation lamellae in the Baraboo quartzite represent a deformation different from that which resulted in the macrofoliation.

On the other hand, the *a*-, *b*-, and *c*-fabric directions so determined could represent the local strain axes of the deformation, independent of the discontinuous mass movement along bedding surfaces. Strata containing axial-plane foliation were initially regions of weakness, and it is along these surfaces (for a wide range in the orientation of the applied forces) that the majority of the mass transport occurs. Referring again to Figure 14, we see that most of the *a–b*-fabric planes are subperpendicular or subparallel to the bedding and the foliation surfaces. If we think of the whole structure as having been formed in one deformation, the regions in which the *a-b* fabric is subperpendicular to the bedding and foliation are regions of approximately horizontal compression parallel to the bedding and foliation. Conversely, regions in which the *a–b*-fabric plane is subparallel to the bedding and the foliation are regions of approximately horizontal stretching parallel to the bedding and the foliation. This latter example is the usual concept of the orientation of the strain in a simple fold. The above-described regions of

compression and stretching in the bedding and/or foliation plane can be thought of as regions of convergence and divergence perpendicular to the principal direction of movement. This phenomenon is recognized in ocean currents and atmospheric currents. Koenigsberger and Hungerer,[27] studying both elastic and plastic bodies, found that in the bending of prisms a secondary fold develops, with a larger radius of curvature than the primary fold and at right angles to the primary fold. These secondary folds have long been recognized by the structural geologists and are called "saddle folds" or "cross-folds." The orientation of the fabric axes determined from the deformation lamellae, therefore, could possibly have resulted from the secondary stresses produced from a series of initially east-northeast folds.

The writer has no preference for either of the above two interpretations. The field problem is always tremendously complicated, and it is almost pure fantasy to interpret a complicated structure in simple terms when we are not even able to interpret a simple structure in complicated terms.

2. In addition to the pattern of the fabric axes determined from the deformation lamellae, the following observations point toward an interpretation of structure that would necessitate more than one period of deformation:

a) Two or more periods of quartz veining are separated by at least one period of deformation.

b) Shear-joint surfaces or perhaps shear fractures in the western half of the region tend to be subparallel to the strike of the *a*–*b*-fabric plane determined from the deformation lamellae.

c) There is chevron folding, that is, buckling, of the axial-plane foliation.

3. A study of a series of samples from the eastern half of the region (in particular, from Sample 53 eastward, in the north range) indicates that the intensity of the deformation has increased toward the nose of the synclinorium. This is evidenced by the following observations:

a) Decrease in the percentage of grains showing original boundaries

b) Increase in the total angular variation in the position of the optic axis in individual grains

c) Increase in the number of grains showing discontinuities in the undulatory extinction

d) Increase in the number of grains which have been broken down into smaller units

e) Increase in the degree of preferred orientation of the optic axes

It is again emphasized that comparisons among samples can be made only when all the samples are of the same composition, that is, in this example, the amount of admixed sericite relative to the quartz. Sericite is a good lubricant, and, when admixed in sufficient quantities to the quartz grains, it will account for the majority of the change in shape of the mixture during deformation.

4. This study has convinced the author that structural petrology needs a theoretical foundation based on experimental deformation of solids and consistent with the physical and chemical concepts of the structure of matter.

[27] A discussion of their work is given by H. W. Fairbairn, *Structural Petrology of Deformed Rocks* (Cambridge, Mass.: Addison-Wesley Press, 1942), pp. 78 and 96.

STRUCTURAL PETROLOGY OF THE PINE CREEK AREA DICKINSON COUNTY, MICHIGAN

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ABSTRACT

The structures of Archean granite and gneiss and overlying Lower Huronian sedimentary formations in the vicinity of Pine Creek in the northern part of the Menominee iron-bearing district of Michigan are compared in this study. Foliation of the gneiss is distinctly unconformable with bedding of the sediments. Faults, joints, and microstructures are similarly developed in the two rock types. During the Killarney orogeny these Archean and Lower Huronian rocks were upturned into a southwest-facing monocline, pushed back at its ends along northeast-trending tear faults. The petrofabrics of both the granitic and the sedimentary rocks plot into horizontal *A–C* girdles and are presumably the result of rotation about a vertical axis associated with the faulting.

INTRODUCTION

The Pine Creek area of southern Dickinson County, Michigan, constitutes the northern portion of the Menominee iron-bearing district. The area occupies a belt 16 miles long and 1–2 miles wide, which trends northwest, generally adjacent to the course of Pine Creek. Included in this belt is the southwestern margin of a complex of granites and gneisses and an almost continuous strip of strongly metamorphosed sedimentary rocks that flank the complex.

In 1904 W. S. Bayley[1] placed these sedimentary rocks in the Lower Huronian and cited occurrences of their basal conglomerates truncating foliation, pegmatite veins, and other structures of the granitic rocks of the Pine Creek area as conclusive evidence that the granites are of Archean age. His opinion remained undisputed in print until C. A. Lamey, discussing the possibility of an extension of the post-Huronian Republic granite, expressed doubt as to the age of the conglomerate and stated that the conglomerate proved "merely that some granite was exposed at the time of its formation, but does not prove the age of that granite or whether its occurrence was widespread or local."[2] F. J. Pettijohn[3] has recently restudied the basal conglomerate and has defended the view of Bayley.

This controversy supplied the impetus for the present study. It seemed probable that a detailed examination and a comparison of the structures, especially the petrofabrics, of the adjacent granitic and sedimentary rocks would give decisive information concerning the relative ages of the rock systems. The author, therefore, spent twelve weeks of the summer of 1946 in the Pine Creek area.

GENERAL GEOLOGY

The Archean complex is composed principally of pink gneisses and granites; subordinate to these are gray gneiss, migmatite, and brick-red granite. All phases of the complex are widespread except the red granite, which forms only 5 per cent of the complex and is restricted to small scattered patches in the northwestern part of the area.

[1] "The Menominee Iron-bearing District of Michigan," *U.S. Geol. Surv. Mono. 46* (1904).

[2] "Republic Granite or Basement Complex?" *Jour. Geol.*, Vol. XLV (1937), p. 507.

[3] "Basal Huronian Conglomerates of Menominee and Calumet Districts, Michigan," *Jour. Geol.*, Vol. LI (1943), pp. 387–97.

The gneisses of the Archean complex are of primary igneous origin. Their folia are due to subparallel arrangement of layers of different composition. These folia vary locally in strike and dip; they wrap around xenoliths; and they wedge out and split into several bands. Much of the pink gneiss contains augen-like phenocrysts of potash feldspar, oriented parallel to the trends of folia, and in many outcrops this foliated pink gneiss grades imperceptibly into lineated pencil gneiss.

The upturned Lower Huronian sediments rest unconformably against the Archean complex. The basal part of the Lower Huronian series, termed the "Fern Creek formation" by F. J. Pettijohn,[4] is composed of 300 feet of conglomerate, arkose, graywacke, argillite, and phyllite. The conglomerate, which truncates structures of the Archean complex, was seen at the following localities: (1) at the falls of the Sturgeon River, Section 8, T. 39 N., R. 28 W.; (2) in the northwestern part of Section 8, T. 39 N., R. 28 W.; (3) at the falls of Black Creek, north-central part of Section 7, T. 39 N., R. 28 W.; (4) near Fern Creek, Section 34, T. 40 N., R. 29 W.; and (5) in the south-central part of Section 29, T. 41 N., R. 29 W.

Above the Fern Creek beds is the Sturgeon formation, which is divisible into a lower member of green sericitic quartzite and an upper member of white, vitreous, fine-grained quartzite. The average thickness of the Sturgeon formation is 1,000 feet. This formation grades into Randville dolomite, the upper member of the Lower Huronian system.

The youngest pre-Cambrian rocks of the area are a series of dikes and mineralized quartz veins, which intrude the Archean complex, and the Fern Creek and Sturgeon formations. The dikes are predominately metamorphosed gabbros and diabases, but in the northwestern part of the area occurrences of melasilexite and esmeraldite are noted.

[4] P. 397 of ftn. 3.

PRIMARY STRUCTURAL UNCONFORMITY

Although the existence of a major angular unconformity between the Archean complex and the Lower Huronian formations is proved by contacts of the basal conglomerates with the granitic rocks at well-spaced localities, this relationship is emphasized by comparing the general strike of the Sturgeon quartzite with the general trends of the foliation of the gneiss throughout the area.

The strike of the Sturgeon quartzite, consistently northwest, actually deviates between N. 15° W. and N. 85° W., and averages N. 50° W In the eastern part of the Pine Creek area the foliation planes of the gneiss trend unvaryingly northwest and form an acute angle with the strike of the quartzite. West of Fern Creek, Section 34, T. 40 N., R. 29 W., the trends gradually change to north and even northeast, the angle between the foliation of the gneiss and the strike of the quartzite increasing to approximately 90°.

The discordance of trends of the quartzite and the gneisses is graphically illustrated by Figure 1, a diagram of the poles to bedding planes of the quartzite, and by Figures 2 and 3, diagrams of poles to foliation planes of the gneisses in the eastern and western parts of the area, respectively.

If the marginal granite gneisses were post-Huronian, their foliation, whether resulting from granitization of sediments or marking lines of flow of an intrusive massive, normally would conform to the general trend of the Lower Huronian rocks. Therefore, the hypothesis of post-

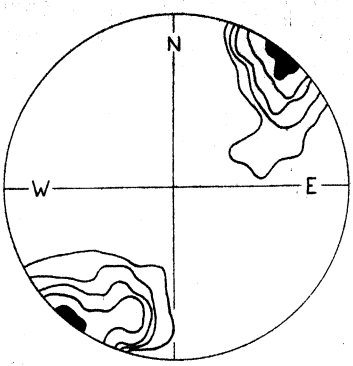

FIG. 1

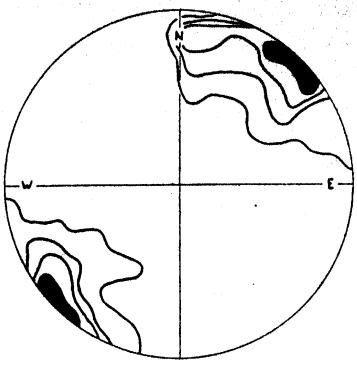

FIG. 2

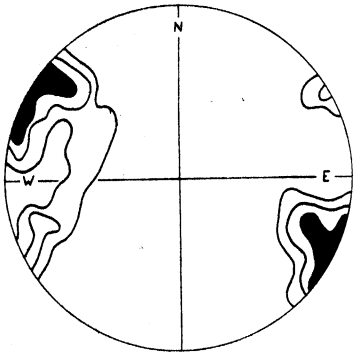

FIG. 3

FIG. 4

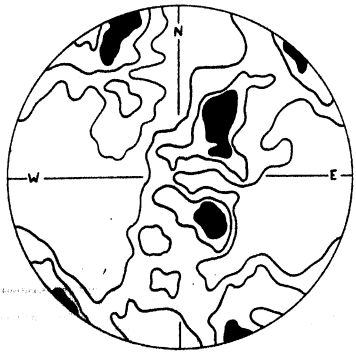

FIG. 5

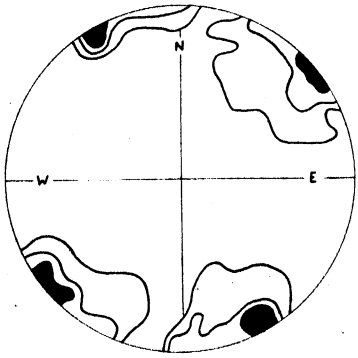

FIG. 6

FIGS. 1–6.—Fig. 1, poles of 119 bedding planes of quartzite west of Section 28, T. 40 N., R. 29 W. (20 per cent maximum). Fig. 2, poles of 50 foliation planes of gneiss east of Section 35, T. 40 N., R. 29 W. (12 per cent maximum). Fig. 3, poles of 70 foliation planes of gneiss west of Section 35, T. 40 N., R. 29 W. (10 per cent maximum). Fig. 4, poles of 76 joints of granitic rocks (4–7 per cent maxima). Fig. 5, poles of 301 joints of quartzite (4 per cent maxima). Fig. 6, poles of 36 joints of basic dikes (10 per cent maxima). All figures are plotted on the lower hemisphere of an equal-area net.

Huronian age of the granite complex is incompatible with the facts.

MAJOR DEFORMATIONAL STRUCTURES

The Lower Huronian strata of the Pine Creek area have been folded and subsequently faulted since their deposition. Both folding and faulting probably were separate phases of the Killarney orogeny.

During the folding the rocks were upturned into a monocline that faced southwest. Dislocated remnants of this monocline are preserved as bluffs of Sturgeon quartzite, which form an arc trending northwest from the center of Section 9, T. 39 N., R. 28 W., to the southwest corner of Section 20, T. 41 N., R. 29 W. The convexity of this line of bluffs is not due to progressive change of strike of the quartzite, however, but to progressive displacement of the quartzite along northeast-trending vertical faults.

The strike of the quartzite, as stated in an earlier paragraph, deviates between N. 15° W. and N. 85° W. and is generally N. 50° W. (Fig. 1). The deviations are local and sporadic and are caused by contortion and rotation of faulted blocks. The dip of the quartzite also has been markedly affected by fault-block rotation. Where faulting is not conspicuous, the beds are vertical or dip steeply southwest. At many places where faulting has been forceful, the beds are overturned and dip northeast, evidently the consequence of tilting of faulted blocks. However, some of the overturning may be the result of underthrusting of quartzite during folding.

Faults of the Pine Creek area are of the tear type. Twenty-one of these faults can be safely inferred and are shown on the structure map, but other faults of inappreciable displacement may exist. Fifteen of these faults are concentrated in the northwestern Pine Creek area. All fifteen strike northeast, their average trend being N. 58° E. Along eleven of the faults the northwest blocks have been displaced northeast through an aggregate distance of more than 3.5 miles. A relatively southwest displacement of northwest blocks along the other four faults totals approximately 1,100 feet.

If the displacements along the fifteen faults are added vectorally and the resultant resolved into east-west and north-south components, the total displacement is 3.5 miles in a direction N. 70° E., the east-west component is 3.2 miles, and the north-south component is 1.2 miles (Fig. 7). The fact that the east-west component has nearly three times the magnitude of the north-south component implies that the forces of faulting were essentially directed from the west.

Related to the faulting of the northwestern area are shear zones strongly developed in both the quartzite and the granite. The rocks in these zones are brecciated or cleaved along closely spaced planes. Some of the zones are found in the immediate vicinities of mappable faults. However, many occur at considerable distances from definite faults, but may themselves be minor faults along which displacement is not prominent.

All except three of seventeen observed shear zones of the quartzite trend between N. 55° E. and east-west. The average trend of all the zones is N. 67° E., differing from the average strike of the faults (N. 58° E.) by only 9° and from the trend of the resultant of the fault movements (N. 70° E.) by only 3°. In several instances shear planes are strongly striated. The pitch of these striations ranges from nearly horizontal (2° NE.) to 40° SW. The pitch of the striations and the average strike of the shear zones

are evidence that the general shear movement was upward at a low angle from west of southwest.

The trends of nineteen observed shear zones in the granite range between N. 15° E. and east-west, and average N. 52° E. This latter figure differs from the average trend of the faults by only 6° and from the average trend of the shear zones in the quartzite by 15°.

Brick-red granite, the only phase of the granitic complex which is nowhere truncated by Lower Huronian basal conglomerate, is intensely sheared in several outcrops. Thus this granite is proved to be older than the faulting, although not necessarily Archean.

At the southeastern end of the Pine Creek area three faults have been mapped. These faults have strikes of N. 44° E., N. 45° E., and N. 60° E., and horizontal displacements of 300, 100, and 100 feet, respectively. In each instance the southeast block has been displaced northeast, the relative movement being opposite to that in the northwestern part of the area.

The central part of the area is but slightly faulted, only three small northeast-trending faults in the vicinity of Fern Creek, Section 34, T. 40 N., R. 29 W., being observed in a distance of 7 miles. Moreover, the quartzite of this area is characteristically sheared parallel to bedding planes. In some outcrops the long dimensions of autoclasts of quartzite are oriented subparallel to the strike of the beds. Therefore, it is assumed that the shearing has resulted from lateral, rather than from vertical, slippage of the bedding planes.

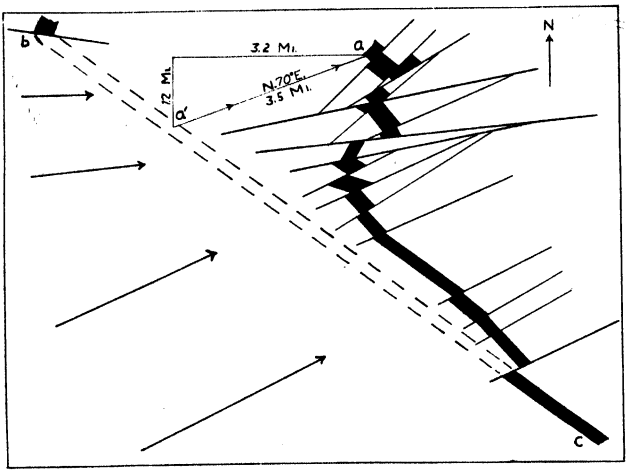

FIG. 7

The faults of the Pine Creek area are not local phenomena but are elements of a regional system. C. E. Dutton and C. A. Lamey have found similar faults in parts of the Menominee iron-bearing district which lie several miles south of the granitic complex. They write:

> Likewise, Although it is recognized that folding is an important factor in the regional distribution of formations, faulting evidently dominates their local distribution. Furthermore, a systematic orientation of faults is apparent, and a well-defined fault pattern comprising (1) a conspicuous set of faults with northeast-south-

west trends, and (2) less conspicuous sets trending (*a*) northwest-southeast, (*b*) nearly north-south, and (*c*) approximately east-west.[5]

The concentration of faults in the northwestern and southeastern parts of the Pine Creek area and the characteristic development of bedding-plane shear in the least faulted central part of the area are best explained by assuming that the Pine Creek monoclinal structure was reinforced by the massive complex of Archean granite. Forces pushing from the southwest against the buttressed central part of the monocline were deflected parallel to the monoclinal front. Although these forces were unable to fault the unyielding granite massive at its center, their deflection was accompanied by lateral slippage of the less competent quartzite along bedding planes. On the other hand, the same forces faulted and pushed back the monoclinal structure at its least resistant ends. The total movement is analogous to that of the front of a glacier thrusting against and moving around a nunatak.

If successive fault blocks in the northwestern part of the Pine Creek area are moved back along their planes, beginning with the fault of Section 20, T. 40 N., R. 29 W., distances equal to their displacements, an approximation of the original front of the Pine Creek monocline can be obtained. The northwesternmost outcrop of quartzite now in the southwest corner of Section 20, T. 41 N., R. 29 W., is moved to a position in the northwest corner of Section 35, T. 41 N., R. 30 W. The same position is obtained if the northwesternmost outcrop is moved a distance equal to, and in a direction opposite to, its total displacement, that is, 3.5 miles in the direction S. 70° W. This position is on the projected strike (N. 53° W.) of unfaulted quartzite, which crops out in Section 28, T. 40 N., R. 29 W., and is less than 3 miles from Sturgeon quartzite near Browning Creek, Section 20, T. 41 N., R. 30 W., which has been described by Pettijohn.[6]

These relationships are pictured in Figure 7. The present and reconstructed positions of the northwesternmost outcrop of quartzite in the Pine Creek area are represented by *a* and *a′*, respectively. The quartzite at Browning Creek is *b*, and the unfaulted quartzite is *c*.

Although the present distance between the Pine Creek and Browning Creek exposures of quartzite is about 6 miles, it is thought that, before faulting, the quartzites at these localities were connected parts of the Pine Creek monocline. This monocline evidently curved northward in the vicinity of Browning Creek, because Pettijohn has determined the average strike of the quartzite at that locality to be N. 35° E.[7]

Two structural features of the Archean complex have resulted from the deformation during the folding of the Pine Creek monocline. The first of these is boudinage structure, occurring in several outcrops in the northwestern part of the area, where lenses of granitic material, 6 inches to 3 feet thick, have been injected into what are now biotite-hornblende schists. These competent lenses dip toward the quartzite and have been broken along their dips into links or boudins surrounded by the more plastic schists. The development of these boudinage structures is attributed to stretching of the granitic lenses as they yielded to forces which folded the quartzite.

A second evidence of deformation of the Archean gneiss is a thrust fault zone

[5] "Geology of the Menominee Range Vicinity of Iron Mountain," *Mich. Dept. Conserv., Geol. Surv. Div. Prog. Rept. 6* (1941), p. 3.

[6] Pp. 395–97 of ftn. 3.

[7] *Ibid.*, p. 395.

in the eastern part of the area. This zone, about ¼ mile north of the margin of the complex, is marked by a sharp reversal in dip of foliation planes of the gneiss and is traceable from Section 9, T. 39 N., R. 28 W., to Section 27, T. 40 N., R. 29 W., a distance of 6 miles (see Map I). Folia of the gneisses in outcrops south of this zone strike unvaryingly northwest and dip southwest, whereas the folia in outcrops north of the zone also strike consistently northwest but dip northeast. Furthermore, northwest strike and northeast dip are characteristic of gneisses throughout the width of the complex north of this zone of dip reversal. Because the northwestern part of the Pine Creek area is so complexly faulted, no integrated zone of dip reversal could be determined there.

Inasmuch as the southwest-dipping gneisses of the margin of the complex, that is, south of the zone of dip reversal, are in direct contact with, and truncated by, the basal conglomerate of the sedimentary series, it is evident that the gneisses at these contacts have been upturned through the same angular distance as the sedimentary rocks. Therefore, the attitudes of the gneisses before deposition of the sediments can be obtained simply by solving the "two-tilt" problem on the stereographic net.[8] This solution involves merely the rotation of a particular plane of foliation about an axis corresponding to the strike of adjacent quartzite and through an angle of such magnitude that the quartzite becomes horizontal.

Folia of gneisses of ten well-spaced outcrops adjacent to quartzite have been rotated in this manner, and their resulting attitudes have been plotted on a map (Fig. 8). To this map have been added typical attitudes of nonrotated folia north of the zone of dip reversal. Finally, the approximate original locations of all folia have been obtained by reconstructing the faulted blocks. The result is a concatenated pattern suggestive of a pluton, the center of which probably was near Fern Creek, Section 34, T. 40 N., R. 29 W., where trends of the foliation change from northwest to northeast.

Therefore, an assumption that the marginal gneisses originally had trends similar to those obtained by rotation, but were subsequently overturned, probably during the formation of the Pine Creek monocline, is indeed credible. Inasmuch as the zone which separates the overturned and nonfolded gneisses is so sharply delimited, it may reasonably be considered a fault zone. This fault probably formed along the axial plane of the monocline and most likely was developed by thrust from the northeast. Such thrusting would have helped to overturn the marginal gneisses.

JOINTING

Similar systems of joints in the granite gneisses and the Sturgeon quartzite are considered evidence that the Archean and Lower Huronian rocks have the same deformational history. The similarity of the joint systems of these rocks is demonstrated by Figures 4 and 5.

One system of joints in the granite consists of a set that strikes N. 45° W. and dips 45° SW., and a weaker set that strikes N. 45° W. and dips about 40° NE. A system of joints in the quartzite includes a set which strikes northwest and dips 45° SW., and a set which strikes northeast and dips 30° NW. If these joints in the granite and quartzite are assumed to be compressional, the bisector of the acute angle of the two sets in

[8] W. H. Bucher, "The Stereographic Projection: A Handy Tool for the Practical Geologist," *Jour. Geol.*, Vol. LII (1944), pp. 203–4.

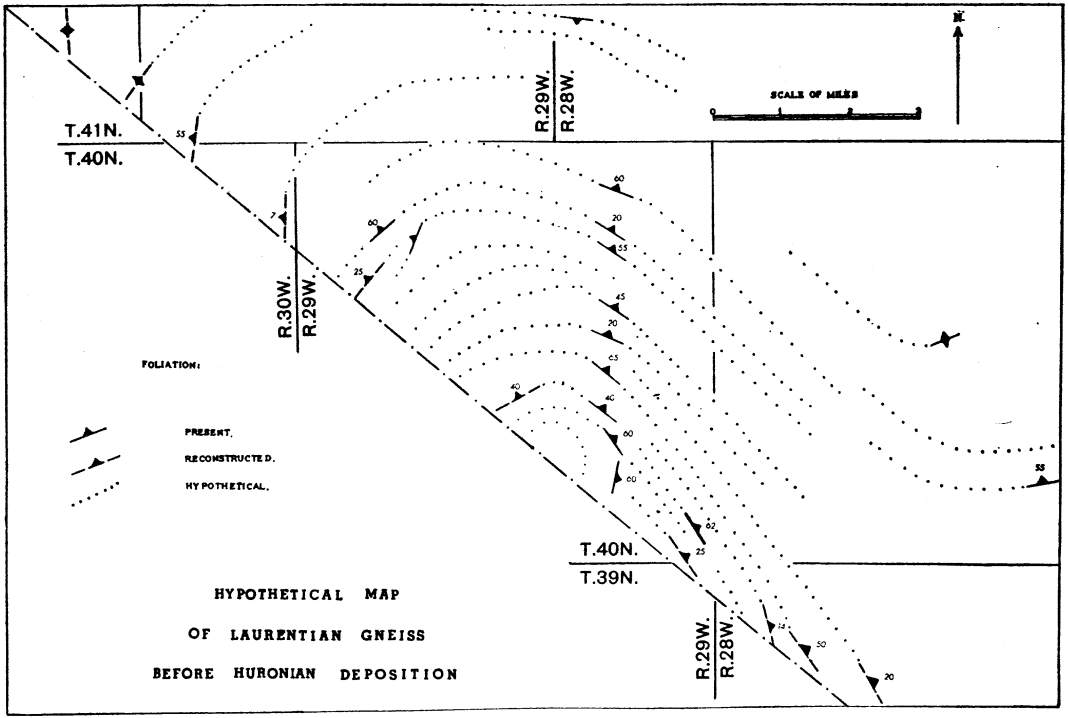
T.41N.
T.40N.
R.30W.
R.29W.
R.29W.
R.28W.
T.40N.
T.39N.
R.29W.
R.28W.
SCALE OF MILES
FOLIATION:
PRESENT.
RECONSTRUCTED.
HYPOTHETICAL.
HYPOTHETICAL MAP
OF LAURENTIAN GNEISS
BEFORE HURONIAN DEPOSITION

FIG. 8

each rock type is equal to the direction of maximum compression. This direction, obtained by use of the stereographic net as outlined by W. H. Bucher,[9] is 15° upward from S. 10°W. if determined by the joints in the quartzite and is nearly horizontal from S. 40° W. if determined by joints in the granite. These directions correspond closely enough to postulate that the conjugate sets of joints of low-angled dip in the quartzite and granite probably resulted from the same forces of deformation. The general direction of the compression which produced these systems of joints is approximately perpendicular to the trend of the Pine Creek monocline. Therefore, the joints may have formed during the folding of this monocline.

A second system of joints, consisting of two vertical sets which trend northeast and northwest, is also equally present in both the granite and the quartzite. The northeast-trending joints, the most conspicuous of the Pine Creek area, locally are the controlling factors of topography. The northwest-trending joints have an average attitude similar to that of the bedding of the quartzite, and many bedding planes are separated by these joints. However, in localities where the bedding differs in strike and dip from the average, these northwest-trending vertical joints are quite distinct from bedding planes.

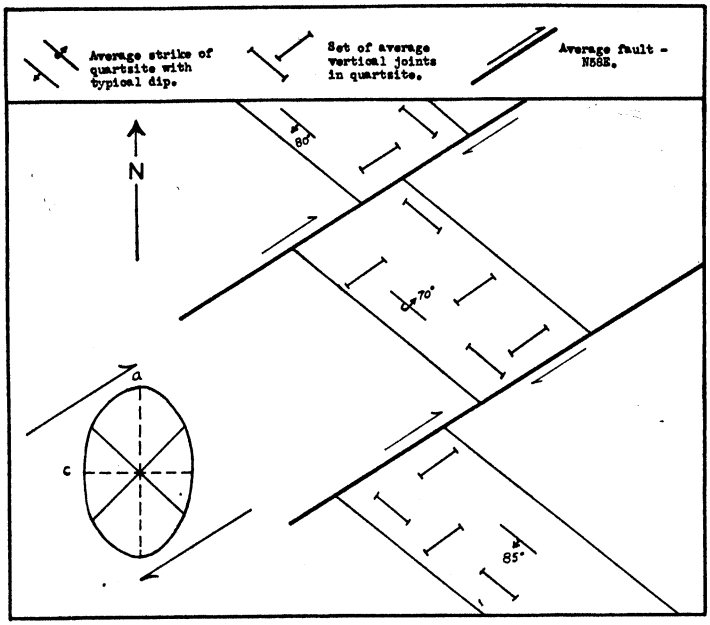

FIG. 9

Both sets of the vertical joints are particularly well shown near zones of faulting, especially in the northwestern part of the area. They may be shear joints related to a couple produced by differential movement of the fault blocks. Figure 9 discloses the possible relationship of these joints to the faulting in the north-

[9] *Ibid.*, p. 196.

western part of the area and to the strain ellipsoid. The strain ellipsoid of the figure indicates that the couple produced an effective component of elongation (*a*) in a north-south direction, and one of compression (*c*) in an east-west direction. The directions of these components of force agree with those obtained by an earlier vectoral analysis of the fault movements of the northwestern area.

The joints of post-Huronian basic dikes of the Pine Creek area belong to a system of vertical sets which correspond in strike to the vertical sets of the granite and quartzite (Fig. 6). Apparently, therefore, the joints of the basic dikes were produced during the faulting. On the other hand, the absence in the dikes of the sets of low-angled joints, common to both the quartzite and the granite gneiss and supposedly related to the folding, is evidence that the dike rocks were intruded either during final stages of the folding or later. Thus the rocks of the granitic complex are not only clearly divorced from, but also definitely older than, the only major features of post-Huronian igneous intrusion. This fact emphasizes the probability of Archean age of this complex.

PETROFABRICS

To supplement the structural data obtained in the field, a petrofabric study has been made of eight samples of the granite and twelve samples of the quartzite. The positions of these samples relative to major structures can be found on the structure map (Map I). The samples of granite and gneiss are designated by the letters A through H, and the quartzite by numerals 1 through 12.

It is not considered necessary to give an elaborate discussion of the theory and procedures of petrofabric analysis. These subjects are treated thoroughly in books by Bruno Sander,[10] Eleanor Knopf and Earl Ingerson,[11] H. W. Fairbairn,[12] M. P. Billings,[13] R. C. Emmons,[14] and in excellent papers by Anna Hietanen[15] and by R. E. Fellows.[16]

METHODS

All samples were oriented in the field relative to some *S*-plane of determined strike and dip, usually a bedding or foliation plane, less frequently a shear plane. Two sections of each specimen were cut for study, one perpendicular to the strike and parallel to the dip of an *S*-plane, the other at right angles to the first, perpendicular to the dip and parallel to the strike.

A study was made of the orientation of the *c*-axes of 200–300 quartz grains in each thin section. For this study a petrographic microscope of the Leitz type and a 4-axis Universal Stage were used. The size of the quartz grains in all samples ranges from less than 0.05 mm. in crush-quartz to more than 1.0 mm. in original grains. Only larger grains were chosen for measurement, as the importance of discrimination in grain selection is becoming increasingly recognized.[17] However, all

[10] *Gefügekunde der Gesteine* (Vienna: Julius Springer, 1930).

[11] "Structural Petrology," *Geol. Soc. Amer. Mem. 6* (1938).

[12] *Structural Petrology of Deformed Rocks* (Cambridge: Addison-Wesley Press, Inc., 1942).

[13] *Structural Geology* (New York: Prentice-Hall, Inc., 1942), pp. 352–55.

[14] "The Universal Stage," *Geol. Soc. Amer. Mem. 8* (1943).

[15] *On the Petrology of Finnish Quartzites* (Helsinki: Government Press, 1938).

[16] "Recrystallization and Flowage in Appalachian Quartzite," *Bull. Geol. Soc. Amer.*, Vol. LIV (1943), pp. 1399–1432.

[17] *Ibid.*, p. 1422; also "The Influence of Grain-Selection on the Meaning of Quartz Diagrams," *Amer. Geophys. Union Trans.* (1945), pp. 653–59.

the large grains were selected at random along five or six well-spaced traverses across the slide. Two of the diagrams are more selective than the others because only those grains showing some degree of undulatory extinction were chosen.

To insure the existence of homogeneity, the grains were plotted in separate groups of 50 or 100, and an inspection made for the tendencies of concentration after each group had been plotted. The plotting was done on the lower hemisphere of an equal-area net of the so-called "Schmidt" type. Concentrations of the *c*-axes were determined by standard center and peripheral counters of 1.0-cm. radius. Counting was spaced by grid distances of 1.0 cm. This was supplemented by closer counting in areas of maxima.

The diagrams which have resulted from the study of sections cut perpendicular to *S*-planes have been rotated into a horizontal plane and are plotted on the structure map. The procedure of rotation of these diagrams was that outlined by Ingerson.[18]

OBSERVATIONS AND INTERPRETATION

The most apparent feature of the fabric diagrams is that they all have peripheral maxima. On the basis of position of the peripheral maxima, the diagrams may be divided into three types.

One type of diagram is characterized by maxima lying between east-west and northeast-southwest, being transverse to any primary structures of the rocks. Diagrams of this type are restricted to samples collected from the southeastern Pine Creek area.

A second distinct type is most common to samples of the northwestern part of the area. Maxima of this type have positions generally north-south. The orientations of these maxima are transverse to bedding planes of the quartzite but usually subparallel to foliation of the gneiss.

To a third type belong a remaining few diagrams of quartzite. The maxima of this type have positions subparallel to bedding planes of the quartzite and may be related merely to residual sedimentary fabric, but more probably to bedding-plane shear. The diagrams of Samples 3, 4, and 5 of the quartzite belong to this type.

Many of the maxima of the diagrams are point maxima, but a few, such as those of Samples 2, 4, and 6 of the sedimentary rocks and Samples A and E of the granite, belong to horizontal girdles. If these horizontal girdles are *A–C* girdles, the directed force must have acted in a nearly horizontal plane about a vertical axis of rotation. Such a force was associated with the faulting. Therefore, a comparison of the fabrics and the faults is desirable.

In order to ascertain just how closely the fabrics of the rocks are related to the faulting, two composite diagrams have been made of samples of the faulted southeastern and northwestern areas. The true value of composite diagrams has been pointed out by F. C. Phillips, who states: "These diagrams, it may be noted, summarize facts of observation. Any conclusions which may be derived from them are unaffected by any of the theoretical considerations of structural petrology."[19]

One diagram (Fig. 10) is constructed from 2,600 grains of Samples 1–5 of the quartzite and Samples A–E of the gneiss. The second diagram (Fig. 11) represents 1,950 grains of Samples 7–12 of the quartzite and Samples F–H of the gran-

[18] Pp. 257–62 of ftn. 11.

[19] "The Micro-fabric of the Moine Schists," *Geol. Mag.*, Vol. LXXXII (1945), p. 206.

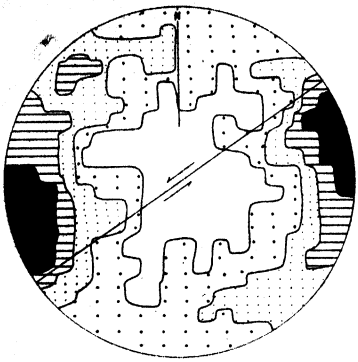

FIG. 10

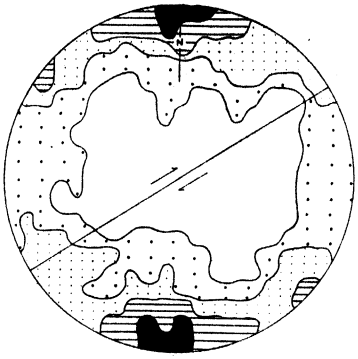

FIG. 11

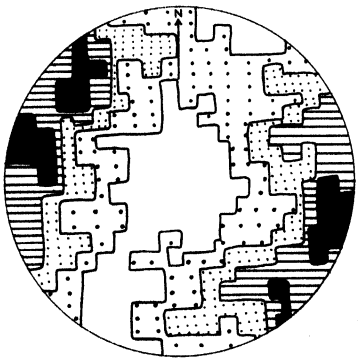

FIG. 12

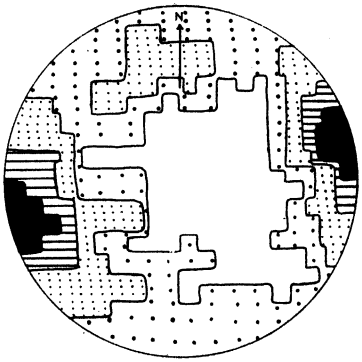

FIG. 13

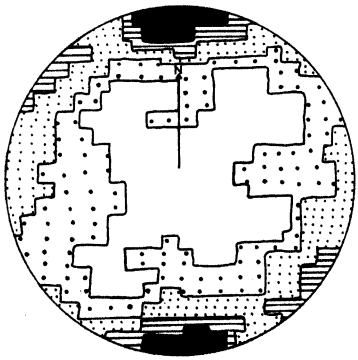

FIG. 14

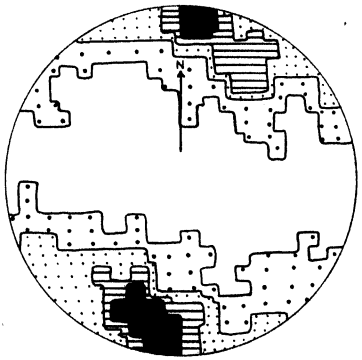

FIG. 15

FIGS. 10–15.—Fig. 10, composite diagram of 2,600 quartz grains of Samples 1–5 and A–E ($\frac{1}{2}$, 1, $1\frac{1}{2}$, 2–$2\frac{1}{2}$ per cent). Southeastern part of Pine Creek area. Fig. 11, composite diagram of 1,950 quartz grains of Samples 7–12 and F–H ($\frac{1}{2}$, 1, 2, 3 per cent). Northwestern part of Pine Creek area. Fig. 12, composite diagram of 1,400 quartz grains of Samples 1–5 of quartzite ($\frac{1}{2}$, 1, $1\frac{1}{2}$, 2 per cent). Southeastern part of Pine Creek area. Fig. 13, composite diagram of 1,200 quartz grains of Samples A–E of gneisses ($\frac{1}{2}$, 1, 2, 3 per cent). Southeastern part of Pine Creek area. Fig. 14, composite diagram of 1,350 quartz grains of Samples 7–12 of quartzite ($\frac{1}{2}$, 1, 2, 3 per cent). Northwestern part of Pine Creek area. Fig. 15, composite diagram of 600 quartz grains of Samples F–H of granite and gneiss ($\frac{1}{2}$, 2, 3, 4 per cent). Northwestern part of Pine Creek area.

ite and gneiss. The average trends of the faults of the areas have been superposed on the diagrams.

The diagram of Figure 10 is a girdle. A peripheral maximum of 2–2½ per cent indicates that a large number of the *c*-axes of the grains are oriented slightly north of east, 25° more to the east than is the average strike of the faults. From the arrows which indicate relative direction of movement of fault blocks, it can be seen that the maximum orientation of the *c*-axes corresponds very well to the direction of elongation of a strain ellipsoid acted upon by a shear couple. The small 1½ per cent submaximum in the northwest quadrant of Figure 10 is possibly due to lateral yielding, through bedding-plane shear, perpendicular to a force from the southwest.

Figure 11 shows that preferred orientation, represented by a 3 per cent peripheral maximum, is north-south. This orientation corresponds to the axis of elongation of a strain ellipsoid acted upon by a couple produced by faulting (Fig. 9); therefore, the maxima are assigned to an *a*-tectonic axis. Fairbairn has cited the development of *a*-maxima in slickensides and states that they "may therefore be considered parallel to the direction of movement."[20] Diagrams of Samples B and H, collected in fault zones, prove the validity of Fairbairn's statement. In the immediate vicinity of the faults the *a*-axis of the strain ellipsoid is tangential to the fault plane. However, it is postulated that, with increasingly greater distances from the fault zone, the *a*-axis of the strain ellipsoid becomes acutely inclined to the strike of the fault. Because most of the rocks studied are located within fault blocks at considerable distances from any fault plane, the maxima of the composite diagrams are developed acutely to the average strike of the faults.

In order to determine whether the fabrics of quartzite are distinctly similar or dissimilar to those of the granitic rocks, separate composite diagrams have been made of the samples of each of these rock types in both the southeastern and the northwestern areas.

The diagrams of the quartzite (Fig. 12) and of the granite (Fig. 13) in the southeastern area are distinctly similar, in that both are girdles with general orientation of quartz axes east-west. A minor difference is that the quartzite diagram has two peripheral maxima, a large one oriented slightly south of east and a smaller one oriented northwest, subparallel to bedding, whereas the diagram of the granitic rocks has but one large peripheral maximum, oriented a few degrees north of east. Preferred orientation subparallel to bedding of the quartzite most likely resulted from bedding-plane shear.

The composite diagrams of samples of the quartzite (Fig. 14) and of samples of granite and gneiss (Fig. 15) of the northwestern Pine Creek area have identical maxima. The only difference between these diagrams is that the maximum of the quartzite diagram is part of a girdle, whereas that of the diagram of the granitic samples is a point maximum.

In summary, it can be stated that the microstructures of rocks of the Pine Creek area are tectonic. Furthermore, all specimens studied have fabrics produced by deformation involving a vertical rather than a horizontal *B*-tectonic axis. Inasmuch as the folding of the Pine Creek area was about a horizontal *B*-axis and the faulting was about a vertical *B*-axis, the rock fabrics are related to the faulting. The general similarity of composite diagrams is adequate proof that

[20] P. 9 of ftn. 12.

the forces involved left the same imprint on both the sedimentary and the granitic rocks.

SUMMARY

Evidences of structure presented in this report are in accord with earlier statements of Bayley and Pettijohn that the complex of granites and gneisses in the Pine Creek area is of Archean age. Trends of foliation of the Archean gneisses are definitely unconformable with the general strike of adjacent upturned Lower Huronian sedimentary rocks. The faults and joints, as well as the microscopic structures of the granitic and sedimentary rocks, are identical.

Also it has been shown that the strata of the Pine Creek area have been deformed at least twice since Archean time, though both deformations probably were separate phases of the Killarney orogeny. During the first phase the Huronian sediments and underlying Archean rocks were folded into a northwest-trending monocline that probably was thrust-faulted along its axial plane. During the second phase of deformation this structure was displaced at its northwestern and southeastern ends by vertical tear faults that trend northeast. The aggregate displacement along faults at the northwestern end has been approximately 3.5 miles. The movement at the southeastern end of the structure was of less magnitude than, and of opposite direction to, that at the northwestern end. The central part of the Pine Creek area was faulted only slightly, but quartzite in this area has been sheared parallel to bedding planes. It is postulated that the faulting and bedding-plane shear have resulted from forces directed from the southwest against the monoclinal structure. These forces were deflected laterally along the central front of the monocline, reinforced by the massive Archean buttress, but were able to displace the monocline at its least resistant ends.

Rock fabrics attest the severity of the fault deformation. The orientation of the *c*-axes of quartz grains in both granitic and sedimentary rocks which crop out in the faulted areas are related to rotational shear about a vertical axis. This shear has been correlated with differential movement of adjacent fault blocks, the components of any one block necessarily being acted upon by a couple.

ACKNOWLEDGMENTS.—The author is grateful to Dr. F. J. Pettijohn for his suggestions and constructive criticisms. He is also appreciative of the able field assistance given by Mr. J. W. Trow and the advice concerning petrofabrics offered by Mr. N. A. Riley.

IGNEOUS METAMORPHISM IN THE OROFINO REGION, IDAHO

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ABSTRACT

A complex series of rocks in the contact-metamorphic zone west of, and adjacent to, the Idaho batholith and underlying the Columbia River basalt is described. The rocks are considered to have been originally chiefly siliceous sediments which have been changed in composition by hydrothermal replacements by solutions from the Idaho batholith to resemble types of quartz diorite and granodiorite. Coarse-grained hornblendite was also formed locally. The over-all textural appearance of these rocks is gneissic, though minor massive and schistose facies occur. Minerals which are considered to have been added to the original siliceous sediments by hydrothermal replacements are biotite, andesine, diopside, uralite, hornblende, scapolite, epidote, zoisite, clinozoisite, allanite, quartz, apatite, sphene, zircon, garnet, magnetite, and sericite.

Ultrabasic rocks in small intrusions, which are described as having occurred somewhat later than the general metamorphism, were altered by autometamorphic, or deuteric, processes.

Introduction

The purpose of this investigation was to determine the nature and origin of the complex, hornblende-rich, gneissic rocks in the contact metamorphic zone of the Idaho batholith, where it is exposed in the lower Clearwater River Canyon, west of Orofino, Idaho. It was thought that these rocks, because of their high hornblende content, the presence of hornblendite dikelike bodies in them, their megascopic appearance, and their position in relation to recognized gneisses formed by granitization, might be somewhat different in character and origin from previously described granitized sediments. A critical microscopic comparison of these rocks with the recognized gneisses formed by granitization in other parts of the region was made, in order to obtain a better insight into their origin. Some small ultra-basic intrusives were studied in less detail to determine their relationship, if any, to the problem.

Field investigation was made during the summers of 1941 and 1942, and the microscopic examination of rock sections was made during the academic years 1941–42 and 1942–43.

The area which is included in this study lies in the western part of Clearwater County and the northeastern part of Nez Perce County, Idaho, and centers about the town of Orofino. That portion of the area which was studied in detail lies along three main streams—the Clearwater River, from about 5 miles south of Orofino to 5 miles west of Peck Station; the North Fork, from Ahsahka to about 5 miles east of Elk Creek; and Orofino Creek, 20 miles eastward from Orofino. The location and boundaries are shown in Figure 1.

General Geologic Setting

The Orofino region is underlain by a thick, complex series of sedimentary rocks, which were metamorphosed by the intrusion of the Idaho batholith during late Mesozoic time.[1] The metamorphic zone is broad and extends at least 30 miles west of the main outcrops of recognized batholithic rocks, indicating that the batholith broadens below the surface and probably does not lie at great depth.

[1] C. P. Ross, "Mesozoic and Tertiary Granitic Rocks in Idaho," *Jour. Geol.*, Vol. XXXVI (1928), pp. 673–93; A. L. Anderson, "The Geology and Mineral Resources of the Region about Orofino, Idaho," *Idaho Bur. Mines and Geol., Pamphlet 34* (1930), p. 18.

The metasedimentary rocks were given the name "Orofino series" and were assigned to the lower Belt series of pre-Cambrian age by A. L. Anderson in 1930,[2] but he now believes that they more probably are Paleozoic or early Mesozoic in age.[3] After the intrusion of the batholith, erosion cut deeply into the metasediments and exposed a large area of the batholith.

During the Miocene epoch, the Columbia River basalt flowed over this irregular erosion surface, covering the metasediments and the margins of the batholith to a depth of 2,000 feet or more. Since that time, basalt has covered the greater part of the area, and the older rocks are exposed only where the streams have cut deep canyons through the basalt and in a few isolated areas where the old erosion surface stood exceptionally high.

The rocks primarily dealt with in this report are the metasediments which crop out along the Clearwater River, the North Fork, and Orofino Creek. They are mainly gneisses, which vary in composition from hornblende-biotite–quartz diorite to granodiorite. A peculiar phase of the hornblende-rich rocks, known as "hornblendite," occurs as dikelike bodies in the gneiss; and coarsely crystalline, pure to silicated marble layers are intercalated with the gneisses. Small bodies of dunite also occur locally in the gneiss.

Detailed structure of the metasedimentary rocks cannot easily be determined because the extent of the exposures is limited by the overlying basalt and talus from that basalt. However, exposures of small extent are plentiful, and, although the rocks have been altered and appear granitic in texture, the original bedding has not been entirely obliterated. Consequently, fairly accurate dip-and-strike measurements can be obtained. The strike ranges in general from east-west to N. 45° W. but varies from due west to east of north. The dip ranges from 15° to 85° to the north. Almost all the rocks have been intricately broken, but no apparent regular fracture pattern could be worked out.

Abrupt changes along the strike from marble to gneiss suggest that transverse faulting may have occurred.

Petrography

Introduction

The rocks of the area are divided into two main groups, igneous and metasedimentary. The igneous rocks are quartz diorite in the Idaho batholith and dunite in small intrusions into the metasediments. The metasedimentary rocks, except marble, have the composition and the textural appearance of igneous rocks and have been so called. However, the rocks occur in distinct, thick to thin, parallel layers which greatly resemble bedding and which, outside this area,[4] grade into almost pure quartzites, on the one hand, and igneous rocks, on the other. Moreover, microscopic examination shows early quartz occurring in granular, more or less equidimensional, interlocking grains as in quartzites and siliceous shales, which indicates that the present composition and texture of the rock are the result of hydrothermal metamorphism of siliceous sediments. Because the metasedimentary rocks, except the marble, have the composition of igneous rocks, the author has given them igneous rock names. These are, in the order in which they will be described, hornblende–quartz diorite gneiss, hornblendite, biotite–quartz diorite gneiss,

[2] P. 9 of ftn. 1.

[3] Personal communication.

[4] A. L. Anderson, personal communication.

granodiorite, and marble. All these rocks, except the marble, are gradational one into another and have been separated on the basis of evidence furnished by microscopic study of thin sections.

IGNEOUS ROCKS

IDAHO BATHOLITH—QUARTZ DIORITE

A large mass of granitoid rock, commonly known as the "Idaho batholith," lies to the east and south of the Orofino region. This batholith has been described[5] as being composed of "two main facies which differ from one another in age, structure, and composition. One facies is represented in a broad marginal and roof zone of more calcic rock . . . .; the other is a less calcic inner zone. . . . ."

Granitic rocks, believed by the writer to be part of this marginal facies of the batholithic mass and to be genetically related to the metasedimentary rocks of the Orofino region, occur at the southern and eastern edges of the area covered in this report (Fig. 1). The rock at the eastern edge of the area is biotite-bearing quartz diorite and is very much like those of the marginal facies referred to above. The rock to the south is similar also but differs in that it contains a rather high percentage of hornblende and a low percentage of quartz.

Since these rocks have been described in detail by Anderson, a brief summary of his petrographic description is given here as follows:

> The marginal facies is mostly quartz diorite. The rock is moderately dark gray, and the granularity is medium to coarse. The essential minerals are sodic andesine (45–60 per cent), quartz (15–30 per cent), and biotite and hornblende (10–15 per cent). Microcline is usually present in very small amounts but locally forms 15–30 per cent of the rock. Sphene, epidote, apatite, magnetite, zircon, allanite, zoisite, chlorite, muscovite, and sericite also were present.

[5] A. L. Anderson, "Endomorphism of the Idaho Batholith," *Bull. Geol. Soc. Amer.*, Vol. LIII (1942), pp. 1102–7.

ULTRA-BASIC ROCK—DUNITE

Field relationships.—Small, irregular bodies of dunite occur in the metasedimentary gneisses. The known areas of outcrop are so small that it is impossible to show them accurately on the accompanying geologic map (Fig. 1). The three occurrences studied are about 3 miles southeast of Teakean, where exposures occur along an area about 10 feet wide an 100 feet long; about 2 miles northwest of Orofino, on the northeast side of the Clearwater River, where a body 10–15 feet wide and 20–25 feet long is exposed; and about 14 miles east of Orofino in a railroad cut, where a dike 10–15 feet wide crops out for ½ mile, disappearing at each end under basalt and taulus. The dunite cuts across the gneisses at varying angles. Near Teakean, where the best exposures were found, the gneisses strike N. 80° W. and dip 75° northeast, whereas the dunite strikes N. 60° W.

Megascopic description.—The dunite is black to green if relatively fresh but becomes greenish-gray to light gray on alteration. It is massive and fine- to medium-grained. Partially altered rock is flecked with areas of fibrous and foliated minerals, which increase in size with progressive alteration until the original structure of the rock is obliterated. The elongated crystals range from fine fibers less than ¼ inch in length to coarse columnar crystals 20 inches long. The foliated material is matted together in aggregates, and therefore the size of individual crystals cannot be determined. Recognizable mineral constituents are olivine, amphibole, and talc, which occur in variable amounts, depending on the stage of alteration attained.

Microscopic description.—The granularity of the fresh rock is generally fine to medium, and the grains are equidimensional, giving the rock its massive structure. The altered rock is fine- to coarse-grained, and the proportion of fibrous to more or less equidimensional minerals varies considerably. Olivine, the chief mineral constituent of this rock, ranges from 100 to less than 5 per cent of the minerals present. The other mineral constituents are amphibole, chlorite, talc, serpentine, quartz, apatite, magnetite, and carbonate.

Olivine occurs in fractured grains ranging from less than 0.25 up to 5 mm. in diameter. The olivine is locally altered to a deep-brown mineral, with high relief and aggregate structure, which was identified as iddingsite. Amphibole, talc, and serpentine penetrate the olivine and isolate grains and parts of grains, accentuating the fineness of the granularity of the rock.

Iddingsite occurs in the dunite east of Orofino. It is dark-brown in color, and some grains show faint pleochroism; it has high relief and displays an aggregate structure. This mineral is an alteration of the olivine and takes the form of the olivine grains when alteration is complete. The change begins along cracks, where the iddingsite has a somewhat fibrous habit. Elsewhere it is fibrolamellar. The writer is not certain of the paragenetic position of the iddingsite. It replaces the olivine and seems to be penetrated by the other minerals as stated. However, it is recognized that the iddingsite may mold itself around the other minerals, as it replaces the olivine for which it seems to have a specific affinity. The water content of iddingsite favors the latter interpretation.

Amphibole occurs as coarse prismatic crystals and as fine prismatic to fibrous crystals, which grow in matted aggregates. The optical properties conform to those of both tremolite and anthophyllite. The grains are colorless and show moderate birefringence. The extinction is parallel or inclined as much as 18°. The crystals are length-slow and display both positive and negative interference figures with large axial angles. The fine material conforms more closely to anthophyllite, which can have inclined extinction in sections that cut all three crystallographic axes. Age relationships between the two are indistinct, but the fibrous material seems to cut across the coarser grains, which conform more closely in optical properties to tremolite. Both seem to pierce olivine and iddingsite, and both appear to be replaced by talc and serpentine, which occur as thin finger-like projections along cleavage directions and fractures in the amphibole.

Colorless to pale-green chlorite occurs in irregular to sheaflike masses. It seems to grow against, and to send minute stringers into, amphibole. Opposite relationships of chlorite to talc are indicated. However, because of the habits of chlorite and talc and because they are probably near the same age, their paragenetic relationships are obscure. Furthermore, the growth of serpentine between chlorite and talc tends to destroy the boundaries between these two minerals.

Talc occurs as very fine aggregates and as flakes up to 1.5 mm. across. It penetrates olivine, amphibole, and possibly chlorite. Serpentine has the appearance of eating into the larger flakes of talc, in places isolating parts which remain in optical continuity with one another. It also sends fine tongues into some smaller flakes of talc and entirely engulfs others.

Long, fibrolamellar crystals of serpentine form in cracks in the olivine and in places surround and isolate parts of oli-

vine grains. In this process fine projections of serpentine penetrate into the olivine, perpendicular to these cracks. Amphibole, chlorite, and talc also are attacked along cleavage directions and elsewhere by serpentine, which penetrates grains and isolates parts of grains.

Quartz, apatite, magnetite, and carbonate form a minor part of the rock, and their detailed paragenesis has not been completely established. Quartz is found as veinlets in cracks or fillings between the other minerals. Magnetite occurs as fine grains or as dust in masses of chlorite, the formation of dust accompanying the development of the chlorite. Apatite and carbonate are rare.

METASEDIMENTARY ROCKS

HORNBLENDE–QUARTZ DIORITE GNEISS

Field relationships.—Hornblende–quartz diorite gneiss crops out along the Clearwater River almost continuously from near Ahsahka to about 4 miles west of Peck Station, where it disappears beneath basalt (Fig. 1). Quartz veins and several hornblendite bodies cut across this gneiss. Pegmatite dikes of composition similar to the gneiss are abundant. The boundaries of the hornblende–quartz diorite gneiss are gradational with those of the granodiorite gneiss, the biotite-quartz gneiss, and the quartz diorite of the Idaho batholith.

Megascopic description.—This rock is medium to dark gray. The granularity is medium to coarse, but fine-grained schistose phases occur locally. Rather distinct gneissic structure is noticeable in outcrops but is not so apparent in hand specimens. Minerals which can be identified megascopically are feldspar, hornblende, biotite, quartz, garnet, and pyrite. Feldspar is generally less abundant in the schistose phases than in the gneiss proper and decreases in amount in the gneiss near the west end of the area. Hornblende increases in amount near the hornblendite dikes and in the rocks west of Peck Station.

Microscopic description.—The hornblende–quartz diorite gneiss is inequigranular-hypidiomorphic. Most of the grains are embayed and penetrated by adjoining minerals, and some large grains tend to engulf smaller ones. Gneissic structure is pronounced to weak, depending on the abundance of feldspar and hornblende porphyroblasts, which tend to grow across and destroy this structure.

Quartz, biotite, plagioclase, and hornblende constitute the major part of this rock. Allanite, epidote, zoisite, clinozoisite, chlorite, sericite, apatite, sphene, zircon, garnet, pyrite, and magnetite occur in variable, but generally minor, amounts.

Quartz occurs in two different habits. The first of these is in small, irregular, more or less equidimensional, interlocking grains, much like that in quartzites and siliceous shales, and is interpreted as remnants of siliceous sediments (Pl. I, figs. 2 and 3).

Biotite also shows dual relationships to the other minerals. Irregular flakes of earlier biotite show rather good alignment. These flakes may lie across several grains, occur between grains, or pierce grains of the early quartz. Because of the relationship of biotite to early quartz and because of field evidence to be discussed under "Petrogenesis," it is considered to be a later mineral than this quartz. The alignment of the biotite is the main cause of the gneissic structure of the rock.

Small, blocky to irregular grains of twinned plagioclase with the composi-

tion of soda andesine near $Ab_{70}An_{30}$ occur in the quartz mosaics and differ in habit and in relationships from andesine of a later stage. Prongs of the irregular grains grow between early quartz and early biotite and penetrate and encircle grains of these minerals, which may become entirely engulfed in the andesine (Pl. I, figs. 3 and 5). Many of the boundaries between this early andesine and early biotite are sharp, but others are ragged, where minute seams of andesine lie along cleavage directions in the biotite. These spatial relationships of early andesine to early biotite and early quartz indicate that it is a later mineral than either of them. Early andesine occurs as inclusions in the second form of andesine, but more details of this relationship will be given below.

Hornblende and biotite constitute 20–40 per cent of the rock in the approximate proportion of 4 to 1. The hornblende shows distinct pleochroism—greenish-gray when the cleavage traces of longitudinal sections are parallel to the vibration direction of the lower nicol and yellowish-green when these traces are perpendicular to this position. Anhedral to subhedral grains as large as 7 mm. in length lie parallel to and across the gneissic structure. Irregular tongues and protuberances of hornblende lie between and penetrate grains of early quartz, early biotite, and early andesine. These three minerals may lie in contact with hornblende or occur in the hornblende entirely isolated from other like grains, giving evidence of the growth of hornblende around them (Pl. I, fig. 1). Late andesine, late biotite, minerals of the epidote group, chlorite, late quartz, apatite, sphene, zircon, and magnetite encroach upon and lie in cracks and cleavages in the hornblende.

Porphyroblastic andesine, also about $Ab_{70}An_{30}$, is more abundant than the fine-grained early andesine described above. Together they constitute 40–60 per cent of the rock. This andesine is in grains as large as 6.5 mm. in length. It shows both albite twinning and pericline twinning. The crystals are subhedral in outline and penetrate irregularly into and between the early quartz, the early andesine, and the early biotite, resulting in all stages of engulfing composites or individual grains of these three minerals. In early stages of growth the late andesine forms lacy, skeletal crystals, the outlines of which are barely discernible in the quartzite (Pl. I, fig. 2). More mature crystals may hold a few isolated grains of early quartz, early biotite, early andesine, and/or hornblende, having replaced all the original intervening material. Porphyroblasts of late andesine, which have very ragged borders because of irregular growth into and between surrounding finer-grained minerals, give the appearance of having been partially crushed; but the diversity of minerals in these fine-grained borders, the absence of cataclastic phenomena, and the presence of replacement relations show that they are not fragments from larger grains (Pl. I, fig. 3). Late andesine also grows into hornblende crystals both across and along cleavages and isolates corroded remnants of hornblende, which remain in optical continuity with the parent-crystal (Pl. I, fig. 4). All these relationships tend to show the replacement character of the late andesine and indicate that it is later than early quartz, early biotite, early andesine, and hornblende. Alteration areas border and occur in many of the andesine grains (Pl. I, fig. 6 and Pl. II, fig. 5). They broaden near and extend along cracks and cleavages, giving the andesine a mottled appearance. Patches of similar material

form, apparently independently of the border areas. However, the border areas and patches are all in optical continuity throughout one andesine crystal, and they all have smaller extinction angles than does the unaltered andesine. Minerals of the epidote group commonly lie inside these altered areas in the andesine. Apparently, cooling solutions penetrated between grains of, and along cracks and cleavages in, this late andesine, changing the andesine to a more sodic plagioclase and using the elements released in the process to form the minerals of the epidote group.

A second-generation biotite differs from the first in habit and penetrates and corrodes the minerals which precede it. Small, ill-formed bodies develop thin, translucent smears and colorations in hornblende; and thin, delicate seams form along cleavage directions and fractures in hornblende and andesine. Larger flakes send fingers between the previously formed minerals. Fine, needle-like projections on the early biotite might be a growth of late biotite, but sufficient evidence for proof of this could not be found.

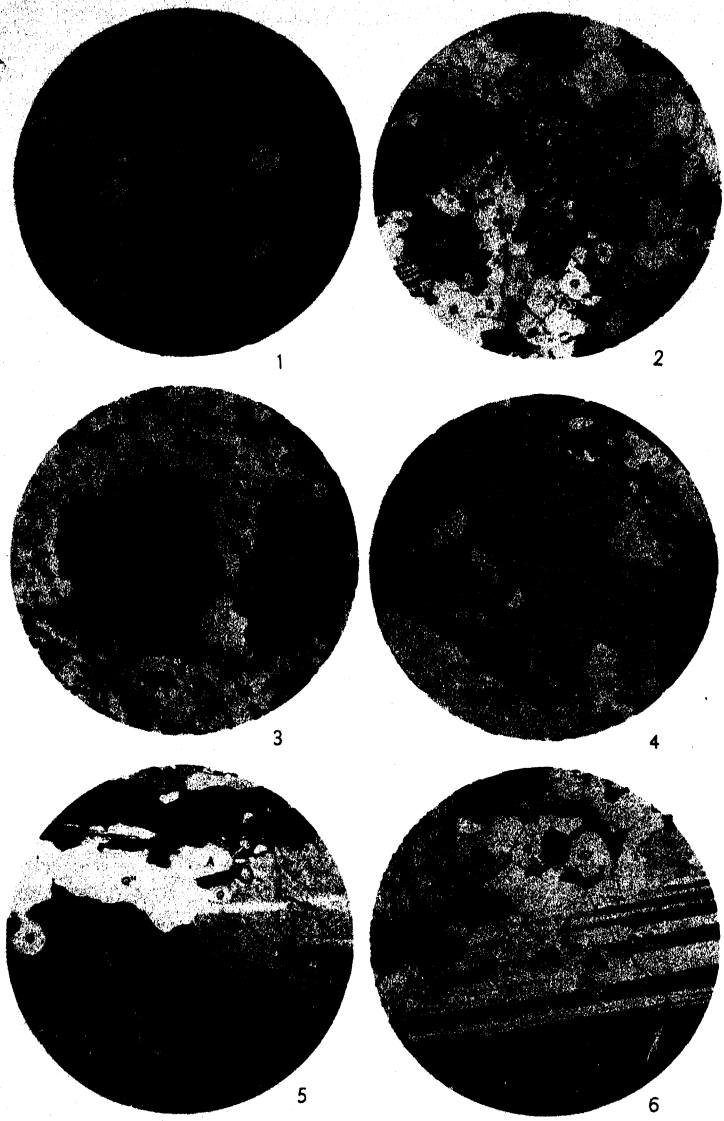
Allanite, epidote, zoisite, and clinozoisite occur in variable amounts, locally forming 10 per cent or more of the rock. All four minerals grow in andesine grains, commonly along cleavage directions and cracks or haphazardly from the sides, and generally lie in or near the alteration areas which were described above (Pl. I, fig. 6, and Pl. II, fig. 5). These minerals also form blocky and irregular grains in hornblende and biotite, apparently growing across the crystals indiscriminately. Allanite commonly occurs as cores around which epidote has grown, but it also occurs by itself. The association of the minerals of the epidote group with the reaction areas in the andesine and the irregular growth in hornblende and biotite crystals suggest that they have formed from the reaction of cooling solutions on andesine and possibly on hornblende and biotite.

Chlorite generally occurs in straight sheaflike, or irregular masses in hornblende and biotite, but it is also present in andesine, in which included biotite has been altered to chlorite. It is scarce, and its paragenetic relationships to minerals other than hornblende and biotite are obscure.

Late quartz differs from the early granular quartz in its habit and spatial relationships to the other minerals. It commonly forms intergranular veinlets and irregular bodies and sends tongues into fractures in adjacent minerals. Large lobate protuberances also extend from the veinlets and irregular bodies and indent andesine and hornblende crystals (Pl. I, fig. 5). Veinlets of late quartz grow irregularly along cleavage planes in hornblende, giving it a motheaten appearance. Early and late quartz together form 5–20 per cent of the rock.

Apatite, sphene, and zircon occur in variable, but minor, amounts. All are closely associated with late quartz and commonly occur as small grains along cracks in the quartz veinlets. In addition, they occur in intergranular spaces and in cracks in other minerals. Apatite also forms blunt, elongate grains, which grow haphazardly into andesine, hornblende, and biotite. Sphene forms large to small irregular grains growing through and around hornblende and other minerals. Zircon occurs as small to minute, subhedral to round, more or less elongate crystals, which are confined chiefly to cracks. There is no direct evidence to show the age relationship between apatite and sphene, but zircon grows along cracks in both these minerals.

PLATE I



Photomicrographs of the hornblende-quartz diorite gneiss. *1*, A hornblende porphyroblast (H) with remnants of early quartz (Q) and early biotite (B) as inclusions. Epidote (E) also is present. Mag. ×50. *2*, A porphyroblast of late andesine (A′) in early stage of growth replacing early quartz (Q) and early andesine (A). Hornblende (H), early biotite (B), epidote (E), and clinozoisite (Cl) also are present. Mag. ×50. *3*, Late andesine porphyroblasts (A′) growing in a mosaic of early quartz (Q), early andesine (A), and early biotite (B). The apparent granulation is a pseudoclastic effect resulting from the replacement of the mosaics by andesine. Mag. ×20. *4*, Late andesine (A′) and late biotite (B′) replacing hornblende (H) which had previously surrounded early quartz (Q), early andesine (A), and early biotite (B). Mag. ×20. *5*, Late andesine (A′) includes a remnant of early andesine (A) in optical continuity with the parent grain, which holds early quartz (Q). Late quartz (Q″) embays both andesines. Mag. ×50. *6*, Epidote (E) surrounded by a reaction area which cuts twinning lamellae in late andesine (A′). Early quartz (Q), hornblende (H), and late biotite (B′) are also shown. Mag. ×50.

1

2

3

4

5

6

Photomicrographs of the granodiorite gneiss and (*5*) of hornblende-quartz diorite gneiss and (*6*) of hornblendite. *1*, Late quartz (Q″) embaying and surrounding late andesine (A′) and microcline (M). Myrmekitic quartz lies in the late andesine in the upper part of the illustration. Mag. ×50. *2*, Microcline (M) isolating late andesine (A′) in which myrmekitic quartz (Q) is growing. Early quartz (Q) is engulfed by late andesine. Late vein quartz (Q″) is also present. Mag. ×50. *3*, Myrmekitic quartz (Q′) in late andesine (A′) which is surrounded and penetrated by microcline (M), which also surrounds but does not replace the myrmekitic quartz as is illustrated in the extreme left part of the figure. Sericite (Se) pierces microcline, late andesine, and myrmekitic quartz. Mag. ×50. *4*, Magnetite (M) surrounds zircon (Z) and apatite (Ap) and penetrates late andesine (A′). Late quartz (Q″) also replaces late andesine. Mag. ×50. *5*, Late andesine (A′) partially altered to epidote (E) and clinozoisite (Cl), which are surrounded by reaction areas (light) in the andesine. Late biotite (B′) also is present. Mag. ×50. *6*, A ragged replacement boundary between late andesine (A′) and hornblende (H) in the hornblendite. Late biotite (B′) is present in both the hornblende and the andesine. The irregular dark line diagonally across the center of the figure is a crack. Mag. ×20.

Subhedral crystals of garnet and pyrite frequently occur together along fairly definite zones parallel to the gneissic structure and where quartz veinlets are common.

Magnetite is in grains of variable size and shape. It grows as blunt to long narrow tongues in interstitial cracks between the other minerals, seemingly widening these cracks during the process. The boundaries of magnetite with almost all the other minerals are generally irregular to ragged. The magnetite also forms narrow, fringelike borders around apatite, sphene, and zircon, somewhat destroying their crystal outlines. Long, irregular tongues and prongs of magnetite grow into and across andesine, hornblende, biotite, and the other minerals, showing the replacement character of the magnetite.

HORNBLENDITE

Field relationships.—Four occurrences of rock, described as hornblendite by A. L. Anderson,[6] are shown on the map (Fig. 1). Hornblendite occurs as dikelike bodies cutting the metasedimentary rocks at various angles to the original bedding. Three of the bodies are less than 20 feet wide, and the other is about 200 feet wide. Numerous other smaller bodies occur in the area, but they are so small that they are not described specifically. All these bodies occur in the hornblende–quartz diorite gneiss. The lengths of the dikes could not be determined because they disappear beneath the basalt talus within short distances. The largest dike crops out on both sides of the Clearwater River about 1 mile west of Ahsahka. The contacts between the dikes and the adjacent gneiss are gradational, and generally the hornblende content of the gneiss increases toward the dikes. Apophyses of hornblendite extend into the gneiss for distances of 3–15 feet from the main bodies.

Megascopic description.—The hornblendite is dark gray to black, very coarse grained, and massive. Ordinarily, the minerals, especially the hornblende, decrease in size toward the margin of the dikes, where they grade into gneisses. The crystals in the small dikes range from less than 1 to 6 inches in length, and in the large dike from 1 to 14 inches in length. The grains of other minerals are not so large as those of the hornblende, but some of the plagioclase crystals are 10–15 mm. across. Stringers and lenses of white feldspar fill cracks in the hornblende. Biotite flakes in discontinuous cross-fractures impart a coarse schiller-like structure to the hornblende.

Microscopic description.—The hornblendite is an extremely coarse-grained, massive rock with hypidiomorphic-granular texture. All the minerals in the hornblendite also occur in the hornblende–quartz diorite gneiss, but the reverse is not true. The relationships of the minerals are the same, but the size of the crystals and the proportions of minerals are different. The minerals of the hornblendite proper are hornblende, biotite, plagioclase, allanite, epidote, chlorite, sericite, quartz, apatite, and magnetite.

Hornblende constitutes 75–100 per cent of the rock. The crystals are very large and have good cleavage. The pleochroism is like that of the hornblende in the adjacent gneiss—greenish-gray when the cleavage traces in longitudinal sections are parallel to the vibration direction of the lower nicol and yellowish-green when these traces are perpendicular to this position. Irregular cracks in this mineral are filled with plagioclase and quartz, probably analogous to the late andesine

[6] "An Occurrence of Giant Hornblendite," *Jour. Geol.*, Vol. XLI (1933), pp. 89–98.

and the late quartz of the hornblende–quartz diorite gneiss, as explained in the following paragraphs. Regular, interrupted cross-fractures at about right angles to the cleavage traces are filled with biotite, probably late biotite, as explained below. Near the margins of the dikelike bodies, hornblende includes remnants of early quartz and early andesine, as described in the hornblende–quartz diorite gneiss. The character of the hornblende, the gradual decrease in hornblende content in the gneiss away from the hornblendite bodies, and the replacement relationships with minerals of the gneiss in the gradational margins of the bodies indicate that the hornblendite represents a localized, special phase of hornblendization in the general process of granitization.

The plagioclase occurs as large crystals, which show very coarse but poorly defined twinning. This plagioclase is soda andesine of about the same composition as the late andesine in the adjacent gneisses. It occurs as veinlets and vug fillings in the hornblende. The boundaries between andesine and hornblende are irregular, because the andesine has made embayments into hornblende and has sent thin seams along the cleavage directions (Pl. II, fig. 6). Isolated grains of hornblende in optical continuity with near-by large crystals are found in the andesine. Near the margins of the dikes, this andesine penetrates and engulfs early quartz, early biotite, and early andesine, as does the late andesine in the gneiss. These relationships tend to show that the andesine in the hornblendite is of the same stage as the late andesine in the hornblende–quartz diorite gneiss.

The habit of the biotite in the hornblendite is similar to that of the second-generation biotite in the hornblende–quartz diorite gneiss. It forms smears, small ill-defined bodies, and large flakes in the hornblende. Other small flakes of biotite fill regular, but interrupted, cross-fractures in the hornblende. In addition, thin blades of biotite occur along cleavage directions and fractures in the andesine (Pl. II, fig. 6). Allanite and epidote appear to have about the same relationships to other minerals that they have in the adjacent gneiss. They grow along fractures and penetrate and lie in andesine, hornblende, and biotite. No evidence to show that epidote is associated with reaction areas in the andesine of this rock was observed. However, the abundance of sericite in the andesine where epidote occurs may have tended to obliterate this evidence. Other relationships are obscure because allanite and epidote are not abundant.

Chlorite grows in irregular and sheaf-like aggregates and is rather abundant, compared to its occurrence in the adjacent gneiss. Definite growth relationships between chlorite and minerals other than hornblende and biotite could not be determined. It grows into the hornblende and biotite along cleavage traces and at the margins of those minerals.

Late quartz occurs much as does similar quartz in the adjacent gneiss but in smaller quantities. It forms veinlets and lobate grains, which penetrate hornblende, andesine, and biotite.

Apatite grows between hornblende and andesine, projecting into both, and occurs in cracks in the andesine. Its association with late quartz is not definite.

Magnetite occurs as rather large irregular grains, which send tonguelike projections along intergranular cracks and pierce adjacent minerals.

BIOTITE–QUARTZ DIORITE GNEISS

Field relationships.—The main outcrop areas of biotite–quartz diorite gneiss are along Orofino Creek, east of the mouth of Whiskey Creek, and along

the North Fork, north of the mouth of Freeman Creek. Two small, isolated areas occur between Orofino and Whiskey Creek (Fig. 1). Quartz veins and pegmatite dikes of composition similar to the gneiss are abundant. The boundaries of the biotite–quartz diorite gneiss are gradational into the granodiorite gneiss and into the quartz diorite of the Idaho batholith. Marble is locally intercalated between layers of the gneiss.

Megascopic description.—This rock is light to moderately gray in color, not so dark as the hornblende–quartz diorite gneiss. The granularity is generally medium to coarse. Gneissic structure is distinct in most outcrops, but some hand specimens appear to be massive. Minerals which can be identified in hand specimens are feldspar, biotite, quartz, hornblende, and garnet. Feldspar, biotite, quartz, and garnet are more abundant than in the hornblende–quartz diorite gneiss.

Microscopic description.—In thin section the biotite–quartz diorite gneiss resembles the hornblende–quartz diorite gneiss in texture, structure, mineral composition, and paragenetic relations. The minerals present are quartz, biotite, andesine, hornblende, allanite, epidote, zoisite, clinozoisite, chlorite, apatite, sphene, zircon, garnet, pyrite, magnetite, and sericite. The chief difference between the two rocks lies in the proportion of minerals. The biotite–quartz diorite gneiss contains a lower percentage of mafic minerals in general, a higher percentage of biotite, and slightly higher percentages of andesine and quartz. These differences are shown in Table 1.

TABLE 1

COMPARATIVE AMOUNTS OF MAJOR MINERALS IN THE GNEISSES AND THE HORNBLENDITE

Name of Rock	Andesine (Per Cent)*	Microcline (Per Cent)*	Quartz (Per Cent)*	Mafic Minerals (Per Cent)*	Biotite-Hornblende Ratios*
Hornblende-quartz diorite gneiss	40–60	.....	5–20	20–40	1:4
Hornblendite	0–25	.....	Less than 5	75–100	1:90 or less
Biotite–quartz diorite gneiss	45–65	.....	10–25	15–30	3:1
Granodiorite gneiss	30–50	8–30	10–25	10–15	3:1

* Approximate.

GRANODIORITE GNEISS

Field relationships.—Outcrops of granodiorite gneiss occur along the Clearwater River from near the mouth of Jim Ford Creek to Ahsahka, along the North Fork from Ahsahka to near the mouth of Freeman Creek, and along Orofino Creek from Orofino to Whiskey Creek (Fig. 1). Quartz veins and pegmatites composed almost entirely of feldspar and quartz are numerous in the granodiorite gneiss, which grades into hornblende–quartz diorite gneiss and biotite–quartz diorite gneiss, with accompanying increases in mafic minerals and plagioclase and a decrease in potash feldspar. Most of the marble of the entire area is interbedded with granodiorite gneiss.

Megascopic description.—The color of the granodiorite gneiss is light to moderate gray, and the granularity is fine to coarse. The rock shows rather distinct light and dark bands, the difference being due to different proportions of feldspar and quartz to biotite and hornblende. Schistose and massive rocks of granodi-

orite composition occur locally in the gneiss but form only a minor portion of the rock as a whole and, consequently, are not shown separately on the map (Fig. 1). In the field the rock is almost indistinguishable from some phases of both the hornblende–quartz diorite gneiss and the biotite–quartz diorite gneiss, the difference being evident only upon microscopic examination.

Microscropic description.—The granularity of the rock is mainly fine to coarse. Some of the larger grains, which are porphyroblastic in origin, are 6–8 mm. in diameter. The actual fineness of grain in places appears to be somewhat accentuated by the manner in which porphyroblasts of plagioclase and microcline penetrate into and between fine-grained earlier minerals, giving the appearance of crushing. None of the major minerals displays euhedral crystal outlines; instead, they are embayed and irregularly penetrated by adjacent minerals.

The texture and the habit of the minerals of this rock are much like those of the hornblende–quartz diorite gneiss and the biotite–quartz diorite gneiss. The differences in the rocks lie mainly in the minerals present and in the proportions of minerals which are common to all three rocks. These differences are shown in Table 1.

The minerals which form the greater part of the granodiorite gneiss are quartz, biotite, hornblende, plagioclase, and microcline. Other mineral constituents which occur in variable and generally minor amounts are pyroxene, uralite, scapolite, allanite, epidote, zoisite, clinozoisite, chlorite, apatite, sphene, zircon, garnet, pyrite, magnetite, sericite, and carbonate.

Early quartz, early biotite, and early andesine in the granodiorite gneiss have properties, habits, and spatial relationships almost identical with those described for the same minerals in the hornblende–quartz diorite gneiss. They occur with the same suite of minerals as is present in the latter rock. Furthermore, they are associated with pyroxene, scapolite, and microcline, which additional associations will be discussed in the descriptions of these minerals.

Pyroxene is rare and occurs only locally. The pyroxene cleavage and the extinction angle of 42°, measured between the cleavage traces in longitudinal section and the *Z*-vibration direction, together with the pale-green color, the rather high relief, and the fairly strong birefringence, show it to be diopside. It forms somewhat blocky to irregular grains, from which prongs have grown between and around the early quartz and the early andesine, completely isolating some of the individual grains. A brown, very finely fibrous mineral, which was identified as uralite, grows along fractures and penetrates into the diopside perpendicular to the fracture surfaces. In some grains small, light-green, pleochroic, spearhead-shaped crystals seem to develop from the uralite and to form hornblende in the diopside. Other diopside grains are penetrated by well-developed prismatic crystals of hornblende, which do not appear to be associated with the uralite. These relationships suggest that the pyroxene is intermediate in age between early andesine and hornblende.

The hornblende is like that previously described in the other metasedimentary gneisses, except for its relationship to diopside and uralite.

Late andesine, although of about the same composition and physical and optical properties as the late andesine of the hornblende–quartz diorite gneiss, has

undergone, as has the early andesine, certain modifications which seem to be associated with the development of microcline. These changes will be presented with the description of the latter mineral.

The properties and spatial relationships of second-generation biotite are essentially the same as those previously described in the other metasedimentary gneisses. The proportion of biotite to hornblende is greater than that in the hornblende–quartz diorite gneiss but is about the same as that in the biotite–quartz diorite gneiss (see Table 1).

Scapolite occurs in some of the granodiorite gneiss which is adjacent to or near marble. The grains of this mineral are larger than the average grains of the rock. They show low relief, moderate birefringence, good cleavages in two directions at right angles to each other, and generally good uniaxial negative interference figures. Scapolite seems to embay diopside and late andesine crystals and, in turn, is penetrated by vermicular growths of clinozoisite. Age relationships with other minerals are not distinct, but the relationships just described indicate that scapolite is younger than late andesine and older than clinozoisite.

The relative age of the minerals of the epidote group to the other minerals and the properties and habits of all except zoisite are like those previously described in the hornblende–quartz diorite gneiss. Zoisite is especially abundant in rocks near or adjacent to marble layers. It has peculiar bluish-white to gray interference colors, and many grains do not become entirely extinct in any position. It forms irregular, coarse, lattice-like growths, with polysynthetically twinned feldspar. Small thin stringers of zoisite pierce diopside, late andesine, and scapolite.

Microcline forms 8–30 per cent of the rock. It occurs in anhedral grains of variable size, which show the gridiron structure and the undulatory extinction common to the mineral. Microcline penetrates irregularly into and between all the minerals described previously under this rock type and in places surrounds them, leaving isolated remnants of these minerals as islands in microcline. Where microcline replaces andesine, myrmekitic quartz develops in the andesine (Pl. II, figs. 1–3). Myrmekitic growths of quartz occur in all the rocks which contain microcline, and all but a few andesine grains which contain myrmekite are definitely in contact with microcline. Even those few grains are in contact with a mineral identified as feldspar and may be either microcline or orthoclase. Apparently, as the microcline replaced andesine, quartz, in an accompanying process, formed vermiform and rodlike bodies in the andesine but not in the microcline. In some corroded grains of andesine the myrmekitic quartz projects into the microcline (Pl. II, fig. 3). This condition is brought about by continued penetration and corrosion of the andesine by the microcline, which causes the myrmekitic quartz, which is replaced less readily than the andesine, if at all, to become partially or wholly enclosed in the microcline.

In addition to the early quartz and the myrmekitic quartz, there is a third generation of this mineral, which occurs as small, irregular veinlets and lobate grains and which corresponds to the late quartz in the previously described metasedimentary gneisses. This quartz displays the same optical properties and spatial relations as does the late quartz of those rocks (Pl. II, fig. 1).

Apatite, sphene, and zircon are more abundant in this rock than in the hornblende–quartz diorite gneiss. These minerals commonly occur as small grains

along cracks in the late-quartz veinlets, in addition to occurring in intergranular spaces and in cracks in other minerals. Apatite and zircon, especially, are commonly surrounded by fringelike borders of magnetite, which somewhat destroy the crystal outlines of the surrounded minerals (Pl. II, fig. 4).

Garnet and pyrite occur locally and sparingly, being commonly associated with late quartz. Some of the garnet is in large grains, which enclose ragged inclusions of biotite and hornblende. Small garnet and pyrite grains are subhedral.

The occurrence of magnetite is the same as it is in the other metasedimentary rocks (Pl. II, fig. 4), and it is considered to be the last mineral constituent formed, with the possible exception of sericite and carbonate.

MARBLE

Field relationships.—Layers of crystalline limestone, or marble, ranging in thickness from 6 to 600 feet, occur intercalated with the gneisses. Most of these layers are interbedded with the granodiorite gneiss in a limited area around Orofino, extending from near Jim Ford Creek, south of Orofino, to about 1½ miles north of Ahsahka. These layers strike in a general northwest-southeast direction more or less parallel to the Clearwater River and dip mainly to the northeast. Outside the area of granodiorite gneiss, about 11 miles east of Orofino at Lime Mountain, a layer of marble about 600 feet thick is interbedded with biotite–quartz diorite gneiss. The marble at this location has undergone more intense silication than that at the other locations described.

Megascopic description.—The marble is white to bluish-gray, coarsely crystalline, and massive. Chemical analyses[7] show that some of the rock is almost entirely calcium carbonate. However, some specimens studied by the writer effervesce only weakly in lump form with cold dilute hydrochloric acid, indicating a fairly high percentage of magnesia. In some places the rock also contains pyroxene, amphibole, feldspar, sphene, magnetite, pyrite, and graphite, all discernible megascopically.

Microscopic description.—The purer rock is almost 100 per cent calcite and dolomite, but it also contains small amounts of talc, quartz, apatite, magnetite, graphite, and pyrite. The silicated rock contains, in addition, diopside, sphene, and zircon.

The carbonate of this rock shows curved twinning lines, which almost obliterate the cleavage traces. The only definite relation observed between cleavage and twinning was that of dolomite twinning parallel to both the long and the short diagonals of cleavage rhombs, seen in a thin section from a specimen, which in lump form did not effervesce freely in cold dilute hydrochloric acid. However, other specimens did react readily to acid, which fact, with the chemical analyses, shows that they contain a high percentage of calcite. In the silicated marble, isolated grains of carbonate, some of which may be secondary in origin, occur in diopside, tremolite, and plagioclase. Veinlets of secondary carbonate fill cracks in other minerals and in carbonate.

Locally, pyroxene may form as much as 80 per cent of the rock. It occurs in large subhedral crystals, which attain a length of 5 mm. The mineral shows good pyroxene cleavage, grayish-green color, rather high relief, fairly strong birefringence, and extinction angles measured to Z of about 42°, all of which are characteristic of diopside. This mineral grows into and around carbonate, leaving small grains scattered through the diop-

[7] A. L. Anderson, pp. 56 and 58 of ftn. 1.

side crystals. Isolated parts of grains of diopside in optical continuity occur in tremolite, plagioclase, and secondary carbonate.

Tremolite forms only a minor part of the rock and generally occurs in small prismatic crystals with diamond-shaped cross-sections, which penetrate diopside and carbonate and form acicular projections along cracks and cleavage directions in these minerals. These spatial relationships indicate that the tremolite replaces the diopside and some of the carbonate and is, therefore, a later mineral than they. Small, irregular, isolated grains occur in plagioclase and secondary carbonate.

Plagioclase is present in subhedral crystals, which vary in size from less than 0.25 to 5 mm. across. Most of the crystals show good polysynthetic twinning, and some of them are also twinned according to the Carlsbad law. Small grains of plagioclase are engulfed by larger crystals of plagioclase, suggesting two stages of development. The smaller, enclosed grains are calcic oligoclase, about Ab_{72}-An_{28}, and the larger crystals are sodic andesine, about $Ab_{68}An_{32}$. Paragenetic relationships between oligoclase and other minerals are obscure, but tremolite seems to cut across and to grow around small oligoclase grains, whereas andesine embays and encloses carbonate, diopside, and tremolite and forms fine fingers along cracks and cleavage directions in the latter two minerals.

A peculiar mineral, which seems to form by the alteration of plagioclase, was identified as zoisite. This identification was based on the association with andesine, the lack of color, the weak birefringence with grayish-green interference colors, and the high relief relative to the adjacent andesine. Because of the habit of the mineral, it was not possible to determine its cleavage, extinction angle, optic character, or axial angle. It occurs in large, irregular, skeleton-like crystals lying in the plagioclase grains. This relationship indicates that zoisite is younger than andesine and has replaced that mineral.

Chlorite occurs sparingly, mainly as fibrolamellar grains in irregular to sheaf-like masses, which seem to grow into the sides of plagioclase crystals. No relationships between chlorite and tremolite or diopside were noted.

Talc occurs in small quantities. In the high-grade marble it grows along cracks in, and sends minute stringers into, the carbonate grains. Some talc is surrounded by vein quartz in the cracks. In the silicated marble, talc penetrates along cracks and cleavage directions in diopside and tremolite.

Quartz forms veinlets and lobate bodies in and between the minerals described above. These bodies penetrate and embay carbonate, diopside, tremolite, and plagioclase. Quartz is not abundant in any of the marble examined.

Apatite and magnetite occur in minor amounts in all the sections studied. They grow along intergranular cracks and penetrate into the bordering minerals in the manners described in the gneisses. Magnetite appears to be the last mineral formed.

In this rock type, sphene and zircon occur only in the silicated marble. Graphite and pyrite form small crystals in the high-grade marble. All four of these minerals are rare, and their paragenetic relationships are obscure.

Petrogenesis

METASEDIMENTARY ROCKS

The bedding occurring in distinct, thick to thin layers, as in sedimentary rocks; the gradations into quartzites, on the one hand, and into igneous rocks, on

the other; and the mosaics of granular interlocking grains of early quartz indicate that the present gneisses were originally quartzites which have been modified by magmatic solutions. Because of the close proximity of the Idaho batholith, the gradational contacts of the gneisses with the batholith, and the mineralogic similarity between the batholithic rock and the gneisses, it is believed that these solutions came from the Idaho batholith, which is exposed east of the metamorphic zone and which probably lies at rather shallow depths below the metamorphic rocks as well. The sequence of mineral deposition indicates that the temperature at which mineral formation occurred increased as the invaded rocks were heated, reached a maximum, and then decreased normally. The action of these solutions on the siliceous sediments is thought to have produced the following changes:

1. Biotite was formed early in the metamorphic process while the rocks were under the stress from the intrusion of the batholith, which condition resulted in the alignment of the biotite flakes. The biotite is considered to have been added and not to have been formed from original constituents in the sediments. This premise is based on the quartzitic character of the quartz and the replacement nature of the biotite and on observations by A. L. Anderson[8] and A. L. Anderson and V. Hammerand,[9] who found in comparable areas of these rocks that biotite occurs in the contact zones but not in the unaltered quartzites.

2. As the temperature of the rocks was raised and the stress relieved, sodic andesine was deposited, partially replacing early quartz and early biotite.

3. When the temperature reached its maximum, diopside was formed locally. Apparently, this condition did not prevail long, because very little diopside was formed except in some of the marble layers.

4. After this high temperature was reached, the solutions became cooler, and some of the diopside was partially altered to uralite.

5. Hornblende formed at the expense of all the minerals which preceded it. The amount of hornblende deposited varied with the access, the amount, and the composition of the feeding solutions. Locally, copious solutions rich in basic materials invaded the rocks and formed coarse-grained hornblendite. According to A. L. Anderson,[10] these solutions were "rendered highly fluid by dissolved mineralizers."

6. Late porphyroblastic andesine penetrated, molded itself around, and isolated previously formed minerals.

7. A second generation of biotite was formed. The relationship of this biotite to hornblende and late andesine proves that it is different from the early biotite. There is some indication that late biotite forms the fine, wisplike projections on the early biotite flakes. J. L. Gillson,[11] in describing a quartz monzonite in the Pioche District, Nevada, mentions such a late growth on biotite. A. L. Anderson and V. Hammerand[12] and A. L. Anderson[13] describe similar growths which penetrate andesine crystals. Both early and late biotite vary in amount from place to

[8] "Contact Phenomena Associated with the Cassia Batholith, Idaho," *Jour. Geol.*, Vol. XLII (1934), p. 386.

[9] "Contact and Endomorphic Phenomena Associated with a Part of the Idaho Batholith," *Jour. Geol.*, Vol. XLVIII (1940), pp. 566–67.

[10] P. 96 of ftn. 6.

[11] "Petrography of the Pioche District, Lincoln County, Nevada," *U.S. Geol. Surv. Prof. Paper 158-D* (1929), p. 88.

[12] Pp. 571 and 574 of ftn. 9.

[13] P. 1105 of ftn. 5.

place, with variations in the quantity and character of the emanations.

8. Scapolite formed locally, replacing diopside and late-andesine crystals.

9. As the solutions penetrated between andesine crystals and along cracks in the andesine, they altered this mineral to a more sodic plagioclase and minerals of the epidote group. As further alteration occurred, some of these products probably were partially dissolved, transported in solution, and redeposited as minerals of the epidote group not associated with andesine but as replacements in hornblende and biotite. It is also possible that these minerals may in part have been formed as alteration products in hornblende.

10. Microcline was deposited in certain restricted areas, indicating further variations in the composition of the emanations given off by the batholith. This mineral penetrates, lies between, and engulfs those minerals which precede it. As andesine was replaced by the microcline, myrmekitic quartz was developed as an accompanying process in the andesine, but in no other mineral. On continued replacement of the andesine, myrmekitic quartz, which is apparently less or not at all susceptible to replacement by the microcline, became wholly or partially surrounded by the latter mineral.

11. Late quartz formed veinlets and lobate bodies, which filled interstitial spaces and replaced adjacent minerals.

12. Apatite, sphene, and zircon apparently followed late quartz, since they occur in cracks in this mineral. Apatite and sphene, and possibly zircon, also replaced andesine, hornblende, and biotite.

13. Magnetite developed along cracks between and penetrated the previously formed minerals.

14. Sericite was formed in the feldspar and carbonate in many of the minerals.

The habits and relative scarcity of garnet, pyrite, and graphite do not permit accurate determination of their paragenetic positions.

Since all the minerals in the gneisses except original quartz were introduced, the solutions which introduced them were of necessity complex. Furthermore, the solutions probably were not always at the same temperature or of the same composition from place to place or from time to time during the period of metamorphism. Stress conditions disappeared or became minor after the biotite stage.

IGNEOUS ROCKS

ULTRA-BASIC ROCK—DUNITE

The minerals of this rock were formed by deposition from, and subsequent reaction by, a cooling magma and its separating solutions. The steps in this process which produced the rock in its present state were as follows:

1. Olivine was deposited from the cooling magma.

2. As the solutions cooled, the amphiboles replaced the olivine. Tremolite seems to have formed earlier than anthophyllite, but their relationships are not distinct, and there may be an overlapping in time of formation of the two minerals.

3. Chlorite replaced amphibole and formed small, irregular, or sheaflike masses in the amphiboles.

4. Talc formed by replacing olivine and amphibole. It may be later than chlorite, or the two minerals may be partially or wholly contemporaneous; but the habits of the two minerals are so similar and their mutual boundaries have been so destroyed by the growth of serpentine that accurate relationships between them cannot be determined.

5. Iddingsite formed around the periph-

eries and in cracks in the olivine, penetrating from the cracks into the grains. The exact paragenetic position of the iddingsite is uncertain, but its high water content indicates that it belongs close to this position in this series.

6. Rather abundant serpentine formed in most of the rocks, replacing olivine, amphibole, chlorite, talc, and iddingsite.

7. Quartz, apatite, magnetite, and carbonate were deposited late, the quartz forming veinlets between most of the other minerals.

The solutions which caused the changes in the dunite are believed to have been simple, cooling solutions and not the complex ones which were required to form the minerals in the gneisses. This is indicated by the normal sequence of mineral formation from olivine to serpentine, the similarity in chemical composition of the minerals, and the lack of parallelism of these minerals to those in the surrounding gneisses either in variety or in temperature of formation. It is believed, therefore, that the rock was altered by autometamorphic, or dueteric, processes in the sense defined by J. J. Sederholm.[14]

Conclusions

1. These complex, hornblende-rich rocks are substantially the same in character and origin, except for the high hornblende content, as the other gneisses of this general region which were formed by granitization.

2. The evidence that the metamorphism described in this paper was caused by hydrothermal solutions is as follows:

a) The gradations of the gneisses into unquestionable quartzites, on the one hand, and into igneous rocks, on the other

b) The fact that thin-section study shows interlocking grains of quartz, which are evidently remnants of quartzite, replaced by a complex suite of minerals common to igneous rocks

c) The sequence, the over-all chemical complexity, and the replacement relations of the constituent minerals of these rocks

d) The mineralogic similarity between the gneisses and the rocks of the Idaho batholith

e) The exceptionally large size of the hornblende crystals which occur in the hornblendite

3. The variations in composition of all the rocks of this area except dunite—as regards, among other minerals, hornblende, biotite, and microcline—are the result of diversity in the structure of the intruded rocks, of variations in the depth of the batholith below these rocks, and of variations in the composition and the quantity of the solutions given off by the batholith from place to place and from time to time.

4. The myrmekite, in these rocks, is the result of replacement of andesine by microcline, with the accompanying development of vermicular quartz in the andesine.

5. The solutions which caused the granitization were complex, hydrothermal solutions. The temperatures of mineral deposition rose as the intruded rocks became heated and then gradually fell to normal levels. On the other hand, the alteration of the ultra-basic intrusives was caused by simple, cooling solutions by

[14] "On Synantetic Minerals and Related Phenomena," *Comm. Geol. Finlande Bull.* *48* (1916), pp. 4 and 142.

autometamorphic, or deuteric,[15] processes.

6. The granitized rocks were under stress during the early stages of granitization, as indicated by the alignment of the biotite.

7. Because of the differences in the character of the solutions which formed the gneisses and the dunite, the latter rock is thought to be different in age from the gneisses; and, because the dunite was not affected by the granitizing solutions, it may well be younger than the gneisses.

8. The hornblendite is thought to represent a special phase of the granitization process because of the similar mineralogy of the hornblendite to that of the gneisses, differing in proportion rather than in character of minerals, and because of its gradational borders into the gneisses.

9. The extremely coarse texture and the high concentration of basic minerals in the hornblendite lead to the conclusion that it was formed by profuse aqueous solutions.

[15] These two terms are here used in the sense defined by Sederholm (*ibid.*).

ACKNOWLEDGMENTS.—An expression of appreciation is herewith given to the Department of Geology, Cornell University, Ithaca, New York, for financial aid in the preparation of thin sections and for use of laboratory equipment during this investigation. Professors J. D. Burfoot, Jr., and A. L. Anderson, of the Department of Geology, Cornell University, gave considerable assistance in the laboratory and offered constructive criticism of the manuscript.

THE COMPOSITION OF METEORITIC MATTER

II. THE COMPOSITION OF IRON METEORITES AND OF THE METAL PHASE OF STONY METEORITES

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ABSTRACT

The average composition, with respect to iron, nickel, and cobalt, of iron meteorites and the metal of stony meteorites has been determined by statistical study of the analyses of 320 meteorites. The average composition of iron meteorites is: Fe, 90.78 ± 0.26 per cent; Ni, 8.59 ± 0.24 per cent; Co, 0.631 ± 0.019 per cent. The average composition of the metal phase of stony meteorites is: Fe, 88.58 ± 0.55 per cent; Ni, 10.69 ± 0.51 per cent; Co, 0.705 ± 0.056 per cent. It is shown that the higher values of nickel and cobalt in the metal phase of stony meteorites result from a relationship whereby the nickel and cobalt concentrations in the metal phase increase as the metal-phase content of the meteorite decreases.

I. INTRODUCTION

In a recent paper,[1] the authors presented the results of a statistical study of a number of analyses of the silicate phase of stony meteorites. The study has been extended to iron meteorites and to the metal phase of stony meteorites. Averages for the composition of iron meteorites have previously been given by O. C. Farrington[2] and by I. and W. Noddack.[3]

The present study was undertaken for two purposes: first, to determine the accuracy with which one might express the average chemical composition of both iron meteorites and the metal phase of stony meteorites and, second, to determine whether or not significant differences in average chemical composition exist between these substances.

II. METHOD AND RESULTS

From the data available on iron meteorites, 220 analyses were selected in which the analyses for iron, nickel, and cobalt appeared to be reasonably satisfactory. Eleven falls and 209 finds were used, of which 151 have been listed by O. C. Farrington,[4] 33 by C. A. Doelter,[5] and 36 have come from more recent sources. In view of the general inadequacy of the analyses for the minor constituents and also in view of the fact that the minor constituents comprise but a small fraction of the whole, the study embraced only the major constituents—iron, nickel, and cobalt. In all cases the sum of these three elements was adjusted to 100 per cent, and the statistical analysis was then made on that basis.

Frequency curves showing the occurrence of iron, nickel, and cobalt in iron meteorites are shown in the upper half of Figure 1. Abundance averages for these elements, together with standard deviations and precisions,[6] are given in Table 1.

One hundred analyses of stony meteorites were accumulated, over half of them falls, in which the metal phase had

[1] "The Composition of Meteoritic Matter. I. The Composition of the Silicate Phase of Stony Meteorites," *Jour. Geol.*, Vol. LV (1947), pp. 405–11.

[2] "Analyses of Stone Meteorites," *Field Mus. Nat. Hist. Geol.*, Vol. III (1911), p. 212.

[3] Die Häufigkeit der chemischen Elemente," *Naturwiss.*, Vol. XXXV (1930), p. 757.

[4] "Analyses of Iron Meteorites Compiled and Classified," *Field Mus. Nat. Hist. Geol.*, Vol. III (1907), p. 59.

[5] *Handbuch der Mineralchemie*, Vol. III, Part II (1926), pp. 574–607.

[6] See ftn. 1.

been analyzed in an adequate manner. Of the stony meteorites used, 71 are listed in the authors' previous paper,[6] and 29 are from other sources. As with the iron meteorites, only data for iron, nickel, and cobalt were used, the sum of these three constituents being adjusted to 100 per cent.

Frequency curves showing the occurrence of iron, nickel, and cobalt in the metal phase are shown in the lower half of Figure 1. Abundance averages for

TABLE 1

THE RELATIVE ABUNDANCE OF IRON, NICKEL, AND COBALT IN IRON METEORITES AND IN THE METAL PHASE OF STONY METEORITES

	Per Cent Fe	Standard Deviation	Per Cent Ni	Standard Deviation	Per Cent Co	Standard Deviation	Total Cases
Metal phase of stony meteorites	88.58±0.55	5.46	10.69±0.51	5.11	0.705±0.056	0.555	100
Iron meteorites	90.78±0.26	3.84	8.59±0.24	3.59	0.631±0.019	0.277	220
Average of Farrington* for iron meteorites	90.85	†	8.52	†	0.59	†	318
Average of Noddacks‡ for iron meteorites	90.2	†	8.46	†	0.547	†	16

* See ftn. 4. † Not determined. ‡ See ftn. 3.

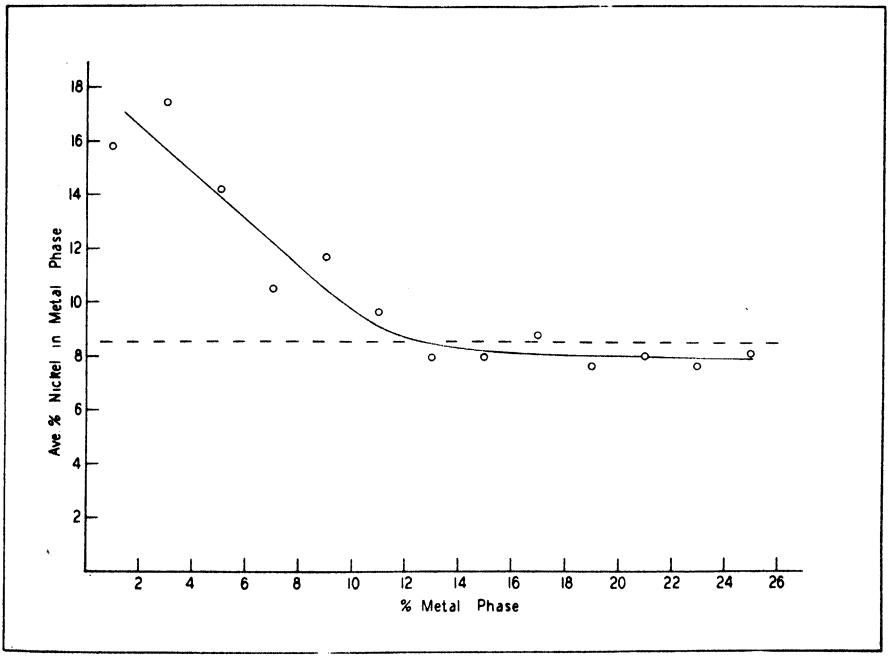

FIG. 2.—The variation of the nickel content in the metal phase of stony meteorites with the metal-phase content in the meteorite. The dotted line represents the average nickel content of iron meteorites. Each point represents the average of between 2 and 19 stony meteorites.

these elements, together with standard deviations and precisions, are given in Table 1.

III. COMPARISON OF RESULTS

A comparison of the figures shown in Table 1 indicates the following facts:

a) The average nickel content of the metal phase of stony meteorites is greater than the average nickel content of iron meteorites.

b) Although the effect is less pronounced, the average cobalt content of the metal phase of stony meteorites is apparently greater than the cobalt content of iron meteorites.

c) As is shown by the standard deviations and by the frequency curves, the spreads in concentration of iron, nickel, and cobalt are greater in the metal phase of stony meteorites than in iron meteorites.

A more detailed study of the data reveals that the higher average nickel and cobalt concentrations associated with the metal phase of stony meteorites results from a pronounced relationship between the nickel concentration in the metal phase and the total metal-phase concentration in the meteorite. If one arranges a statistical sampling of stony meteorites in the order of increasing metal-phase concentration and averages the nickel concentrations over specified intervals of metal-phase concentrations, the curve shown in Figure 2 is obtained. It can be seen from Figure 2 that, as the concentration of metal phase associated with meteorites decreases, the nickel concentration in the phase increases to a value that is approximately double the average nickel concentration in iron meteorites, represented by the broken line. The thirteen points defining the curve represent averages calculated from 96 stony meteorites, of which 60 had a metal-phase content of less than 10 per cent.

The rapid rise in nickel concentration with decreasing metal-phase concentration begins at a metal-phase concentration of approximately 13 per cent. At metal-phase concentrations greater than 13 per cent, the nickel concentration levels off to a value slightly lower than, but nevertheless very close to, the average nickel concentration in iron meteorites.

These latter observations substantiate conclusions to the same effect drawn previously by G. T. Prior.[7]

[7] "On the Genetic Relationships and Classification of Meteorites," *Min. Mag.*, Vol. XVIII (1916), p. 26; and "The Classification of Meteorites," *ibid.*, Vol. XIX (1920), p. 54.

UPRIGHT TRUNKS OF *NEOCALAMITES* FROM THE UPPER TRIASSIC OF WESTERN COLORADO

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ABSTRACT

This article describes a hitherto unreported find of *Neocalamites* from the Upper Triassic of western Colorado. The find is of unusual interest because of the large size of the trees and their geologic mode of occurrence. Moreover, the large number of plants exposed indicates the presence of an Upper Triassic forest of considerable size.

INTRODUCTION

Many fossils and something of the woody structure and life-history of the Equisetales are fairly well known from the Devonian to the end of the Paleozoic, when the well-known genus *Calamites* became extinct. But our knowledge of the history of this great group of plants after the end of the Paleozoic is rather meager. That the group survived the Paleozoic is known from scattered finds reported from time to time from the Mesozoic and from the single present genus *Equisetum*. These modern horsetails are the only living representatives of a much greater group of plants that dominated the Carboniferous landscape.

The *Calamites* of the Carboniferous grew to great size. Trunks 30 feet long and 1 foot in diameter have been reported. The number of individuals must have been very great, for their casts and impressions are found in large numbers in Upper Paleozoic rocks.

When we come to the Mesozoic, however, the Equisetales, though still found, are rare, and those that have been reported are usually of small size compared with their Paleozoic predecessors.

The find described here from the Upper Triassic of western Colorado is therefore of more than passing interest, not only because of its rarity but also because of the large size of the fossil trunks. Furthermore, the plants are all still in the vertical position of growth, and consequently have not been compressed or distorted.

These fossils were first discovered by the writer in the spring of 1942, but further study was interrupted by army service until 1946, when another trip was made into the area. The fossil locality is in west-central Colorado, on the south side of the Dolores River and about a mile from the town of Bedrock, Colorado.

Preliminary examination indicates that the forest here discovered is comparable in many respects to that of the celebrated Joggins, Nova Scotia, section of Carboniferous age, where upright trunks of *Calamites* and *Sigillaria* have excited the admiration of geologists since the days of Lyell, Logan, and Dawson. According to Charles Schuchert and C. O. Dunbar,[1] some of the trees at Joggins have a length of 20 feet, and a few are 4 feet in diameter. While these trees from Joggins are much thicker than those described here, it is believed that these Triassic plants were fully as long. In both localities, however, the tops have been broken off, thus making it impossible to

[1] *A Textbook of Geology*, Part II, "Historical Geology" (New York: John Wiley & Sons, Inc., 1933), pp. 249–50.

get a true picture of the actual height. A comparison of the photographs of the Colorado plants with those shown by Schuchert and Dunbar[2] of the Joggins fossils shows a striking similarity.

STRATIGRAPHY AND MODE OF OCCURRENCE OF THE PLANTS

The Triassic formations of the Colorado Plateau province consist of a great series of red sandstones, shales, and conglomerates, mostly nonfossiliferous and probably in large part of continental origin. In Arizona and Utah these "Red Beds" have a threefold division, the Moenkopi, Shinarump, and Chinle. According to J. Gilluly and J. B. Reeside,[3] these strata are 2,232 feet thick in eastern Utah. However, the Triassic rapidly thins eastward into western Colorado; and, in view of this thinning to the east, it is doubtful whether either the Moenkopi or the Shinarump is represented in the fossil locality under discussion.

The Chinle formation of western Colorado, in which the fossils occur, is considered Upper Triassic in age. In this area it has a total thickness of about 300 feet and typically outcrops in fairly steep slopes at the base of the Wingate sandstone. Opinion as to the age of the Wingate sandstone has varied with different writers. Some have assigned it to the Triassic and others to the Jurassic. It is a great, cliff-making sandstone in the plateau country and stands out as a sheer cliff several hundred feet high in the canyons, while the Chinle forms steep slopes below it.

In this area the Chinle formation consists of micaceous red-brown sandstones, maroon sandy shales, and some thin limestone conglomerates. The sandstones are uniformly fine grained, few grains being large enough to be easily recognized. The red sands in which the fossils occur are also of very fine grain, and there is an absence of well-defined bedding planes. Two fossiliferous conglomerate layers, from 12 to 18 inches thick, containing *Unio* and bone fragments were found at the fossil-plant locality, one above the tops of the trees and the other just below the roots. These conglomerates carry limestone pebbles as much as ½ inch in diameter, together with well-rounded quartz and feldspar fragments.

The fossil plants are embedded in this fine-grained sandstone between the two conglomerate layers and about 150 feet below the Wingate sandstone cliff. The trunks of the trees occur as casts, and some measure 12 inches in diameter. All trunks seen were *in situ* and in the vertical position of growth (Pl. I, *A*). They are quite close together. The trunks of five trees were fairly well exposed, and further search revealed the tops of many others projecting above the surface. A 15-foot section of one tree was removed, and some preliminary excavating was done in the hope of finding the roots, but this attempt was abandoned because of lack of time. Considering the 15-foot length of the specimen collected and the fact that the root was not reached and that the top was broken off, it is estimated that these giant Mesozoic Equisetales may have reached a height of well over 50 feet. The casts themselves are fairly well preserved and solid and show quite clearly the parallel grooves formed by the vascular bundles (Pl. II, *A*, *B*). The nodes and leaf scars are also well preserved, but no woody structure of the actual plant has yet been found. Although no leaf impressions were found, enough of the structure of the nodes is

[2] *Ibid.*

[3] "Sedimentary Rocks of the San Rafael Swell and Some Adjacent Areas in Eastern Utah," *U.S. Geo. Surv. Prof. Paper 150-D* (1928), p. 65.

PLATE I

A

B

A. View showing Neocalamites erect in the position of growth

B. Neocalamites *in situ*, western Colorado

PLATE II

Close-up of one trunk showing nodes and grooves made by the vascular bundles

preserved to identify the plants as *Neocalamites*.

L. H. Daugherty[4] has reported the presence of *Neocalamites* in the Upper Triassic of Arizona, but the plants that he describes are considerably smaller than these in Colorado. It is believed that the *Neocalamites* here described are the largest yet reported from the North American Triassic. Possibly these trees had reached their maximum size at that time, judging from the fact that those reported from higher horizons in the Mesozoic are, on the whole, noticeably smaller.

CONDITIONS OF SEDIMENTATION

Perhaps the most striking feature of the fossils is the evidence which they present of very rapid sedimentation. Although these hollow-stemmed plants, as now preserved, are about 20 feet high, yet they were rapidly and effectively buried before they could fall. The absence of coarse sediment and well-defined bedding planes in the strata apparently would indicate deposition under uniform conditions and probably by slow-flowing currents. Yet, if this is true, it is hard to see how slow-flowing currents could bury the plants before they fell. Possibly, sluggish, meandering streams, heavily laden with sediment, broke through natural levees of their channels onto the timbered low lands during the Chinle epoch of sedimentation and thus rapidly buried the trees.

Another problem is how the sands got into the hollow stems to form the casts. Two possibilities are suggested. First, the tops of the trees may have been broken off after rapid sedimentation had buried some 20 feet of their trunks, and sand then had access to the hollow interiors. A second possibility is that they were completely buried before the woody structure of the outside decomposed, thus allowing for the formation of molds, which were subsequently filled with sands. A careful search has revealed no trace of any woody structure; consequently, the writer is inclined to believe that the tops of the plants were broken off and the hollow interior subsequently filled with sand.

ACKNOWLEDGMENTS.—The writer wishes to thank Dr. John B. Reeside, Jr., and Dr. Ralph Brown, both of the United States Geological Survey, for reading the manuscript and making some helpful suggestions.

[4] "The Upper Triassic Flora of Arizona," *Carnegie Inst. Wash. Pub. No. 526* (Washington, D.C., 1941), p. 58.

THE RATE OF ROUNDING OF BEACH BOULDERS

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ABSTRACT

A demonstration is given of the ability of waves of moderate strength to round blocks of hard, fine-grained basalt as much as 2 feet in diameter in less than ten years. Mention is also made of an example of the short distance of travel that may be involved in the rounding of stream pebbles.

A little north of Takapuna Beach, which lies in Hauraki Gulf about 5 miles north of Auckland City, New Zealand, there has been demonstrated within recent years the rapidity with which waves of moderate strength are able to round blocks of hard rock. As quantitative data upon phenomena of this kind are relatively scarce, the writer has ventured to give the following account of the facts.

About twenty years ago the local borough council unwisely allowed the removal of a natural barrier of blocks of basalt which had long protected the shore from the sea waves at a locality indicated by the tip of the arrow on the map (Pl. I*A*). Here a subhorizontal sheet of basaltic lava, from 2 to 3 feet in depth, rests near high-water mark upon an unknown thickness of soft white "pipe clay" (Pl. I*B*), in some places with the intervention of a few inches of volcanic cinders. Where not highly vesicular, as it may be locally, the basalt, which is a hard, fine-grained, olivine-rich variety, is dissected by vertical joints into irregular polygonal columns usually from 18 inches to 2 feet across.

Upon removal of the blocks of basalt that had earlier formed the shore, the waves soon excavated the soft pipe clay from beneath the basalt, thereby undermining the latter and causing its collapse as a relatively unbroken sheet along a strip extending back from the earlier shoreline for about 6–8 yards. It was considerable time before the waves disintegrated this sheet into blocks sufficiently isolated to be subject to rolling one on another, and it is only within the last ten years that such rolling has been an effective process. During that time there has been continued, but slow, collapse of the edge of the sheet of basalt until its maximum recession is now probably about 15 yards, although the writer has no exact data upon this point. Plate II is a typical view (at low tide) of the boulders that strew the shore; many of the larger are still angular, but others of from 1 foot to over 18 inches in diameter are well rounded.

In comparison with exposed shorelines, the locality at which these boulders occur is sheltered, for it is protected from major swell by outlying islands, as is evident from the map (Pl. I*A*). The waters offshore are shallow, for at ½ mile from the shoreline the depth at mean low tide is a little over 3 fathoms; it increases to 5 fathoms at 1 mile and reaches its maximum of from 6 to 7 fathoms at 1½ miles. Although there is open water subject to frequent northeast gales for nearly 50 miles to the northeast, the boulder beach itself is not exposed directly to the waves generated by these gales, being protected by a short, low tongue of unbroken basaltic lava, which is shown in the background of Plate II.

Although waves have been stirring the boulders of this shore more or less effec-

PLATE I

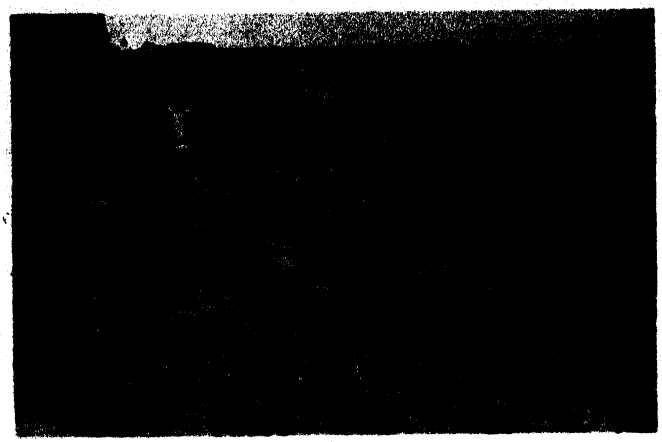
A

B

A.—Map showing Hauraki Gulf near Auckland, New Zealand. The area described is at the tip of the arrow.

B.—Beach boulders and rubble lying on soft white "pipe clay" (foreground) at the shore near Takapuna Auckland, New Zealand.

PLATE II



Boulders at the locality of Fig. 2. The largest spherical boulder measured is 1 ft. 10 in. in diameter.

tively for approximately ten years, it is only within the last four or five years that the result of their work in rolling, and thereby rounding, the large blocks of basalt has become obvious.

An observation pertaining to a related phenomenon was made on the rounding of stream pebbles in the vicinity of Whangarei, North Auckland. At McLeod's Bay, Whangarei Heads, a small stream rises in a sharp-crested range made of coarse-textured andesitic breccia and plunges down the precipitous face of the range for about $\frac{1}{4}$ mile before flowing on a moderate gradient to the shore at a little over $\frac{3}{4}$ mile from the divide. Where exposed by road cuts, its earlier fan-gravels include pebbles up to 2 inches in diameter, of which about 60 per cent show a surprisingly high degree of rounding. The hard, fine-grained andesite of which they are composed accepts a very smooth surface upon attrition, and a very large proportion of the rounded pebbles show perfection both of rounding and of smoothing. Prior to seeing this demonstration of the work of streams, the writer had imagined that far greater length of travel would be required before perfection of rounding could be attained; this misconception may be fairly general.

THE ORIGIN AND IMPORTANCE OF THE RALEIGH GRAPHITE

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ABSTRACT

There are several parallel tilted beds of earthy schistose graphite enclosed in the pre-Cambrian schists and gneisses of Wake County, North Carolina. Two or these beds are large enough to be of commercial interest and therefore were mapped along a 10-mile strike. From their structure and mineraliaztion it is evident that the graphite is part of a series of metamorphosed carbonaceous sediments. These beds will prove to be excellent key horizons for later work in the area.

INTRODUCTION

Schistose earthy graphite occurs in many beds in the pre-Cambrian schists and gneisses of Wake County, North Carolina. These graphite beds are grouped together and discussed in this paper as the "Raleigh graphite." All the exposures of the Raleigh graphite which were mapped lie in a belt 10 miles long and 2 miles wide. The northern end of the belt is at the town of Six Forks, nearly 10 miles north of Raleigh. How far south it extends has never been satisfactorily determined. Its southern end may be more than 18 miles southwest of Raleigh.[1] This area lies entirely in the Raleigh quadrangle of the United States Geologic Survey (Fig. 1).

DESCRIPTION OF THE GRAPHITE

The graphite is schistose with well-developed foliation. There is a variation in the appearance of this schistosity which is a direct reflection of the percentage of graphite in the rock. At most localities the schist averages about 4 per cent graphite. In this graphite-lean rock the foliation is extremely well developed; the schist is quite micaceous, light- to dark-gray, and quite hard. It will scratch a piece of paper rather than mark it, although black graphite may be rubbed from a hand specimen with the fingers. Locally, the graphite content of the schist may reach 35 per cent. If it is above 12 per cent, the schist is more massive, almost slaty, black, and very soft. Graphite of this type will mark a piece of paper easily. There is still a visible quantity of mica in the schist, although it may not show on the outside of a weathered lump. Specimens taken from old mine dumps appear to be pure graphite from the outside, as the mica has been washed off. A much better picture is obtained from the inside of these lumps, where the mica shows as minute, sparkling, contrasting flakes.

Quartz is a major constituent of both types of graphite schist and composes from 40 to 85 per cent of the rock's weight. Some of this quartz is plainly seen in veins and stringers. Most of it, however, is in a finely divided state and consequently difficult to see. It is also partly hidden by graphite, which has stained the minute cracks in its surface. Both types of schist, when ground to −200 mesh, almost invariably show quartz and graphite in every fragment. Garnet and kyanite occur in the graphite in a few places but are not common. These two minerals are localized in zones and are later than either the graphite or the schistosity.

[1] Ebenezer Emmons, *Geological Report of the Midland Counties of North Carolina* (New York and Raleigh, 1856), p. 221.

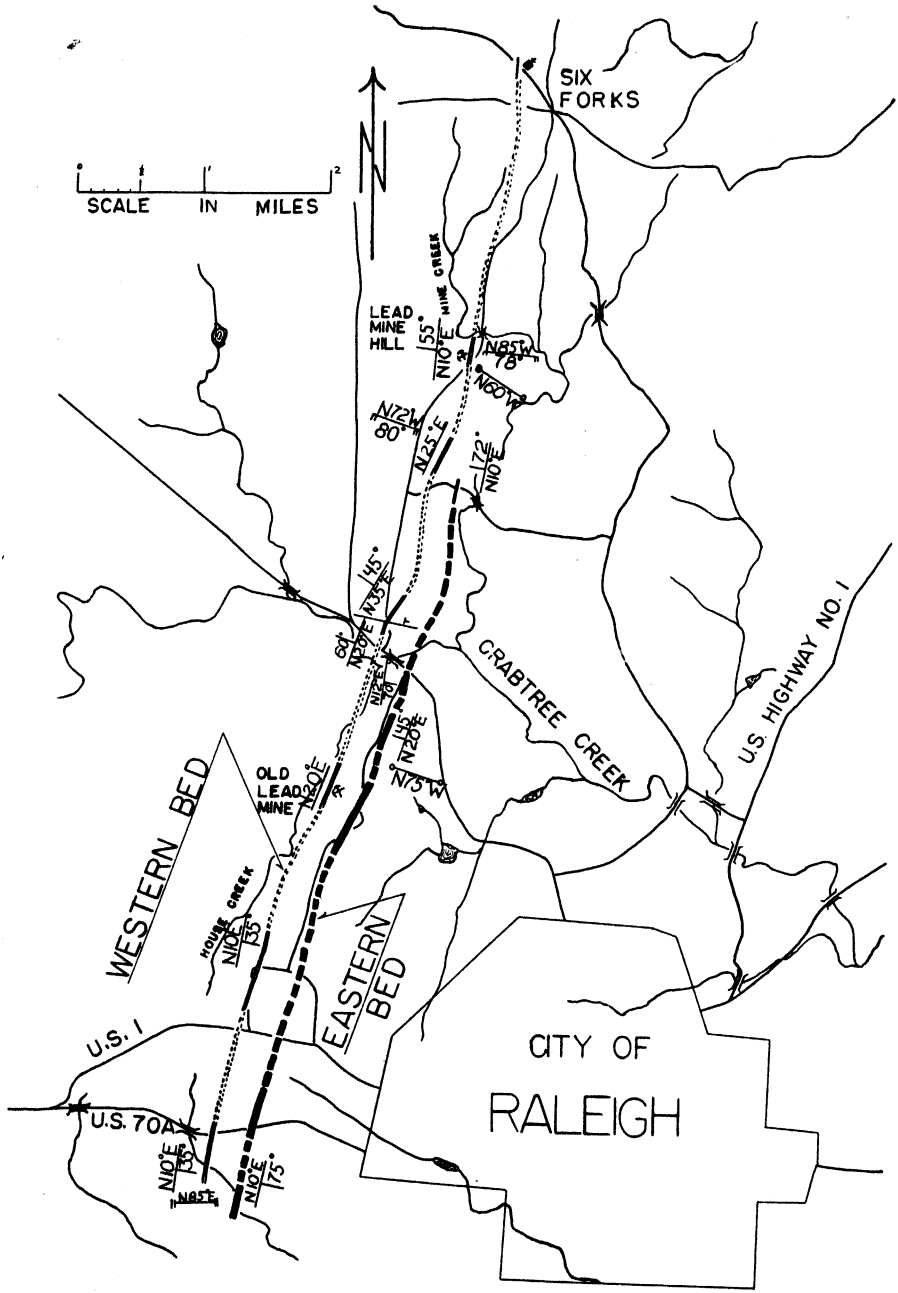

FIG. 1.—Geologic map of the Raleigh graphite

OCCURRENCE IN BEDS

It is difficult to select a term to describe the occurrence of the graphite which does not also predicate an origin. The term "bed" was selected, as it gives a clear picture of the interbedded appearance of the graphite layers and the schistosity of the enclosing gneisses and schists. Justification for this special usage is given in the section on the origin of the graphite.

Only two of the beds of graphite schist are wide enough to be considered economically important. These contain all the old mines and prospect holes, and they are the only beds mapped throughout the 10 miles. Both beds strike about N. 12° E. and are parallel to the schistosity of the country rock. A few miles north of Raleigh they curve slighly to the northeast. This change of strike direction is attributed to rotational faulting rather than to a simple bending of the regional schistosity.

There is but one fault plane exposed; however, other strike changes and outcrop locations indicate that there may be at least three more. This fault plane is clearly exposed in a road cut, striking N. 80° W. It is practically vertical. On the south side the graphite strikes N. 20° E. and dips 60° W. On the north side the same band strikes N. 35° E. and dips 45° W. There has been a torsional rotation of the graphite bed through an arc of 15°. The strike direction also changed 15°.

The geometry of such a fault is complicated beyond simple analysis by the degree of flexure, which may have been achieved before rupture took place and the fault plane itself developed. If a piece of paper is held so that it dips 60°, and then one end is twisted to give a change of dip to 45°, an unruptured flexure is formed. The strike changes along a horizontal section from the original strike at one end to a new strike at the other. Experiment with rigid and plastic models indicates that rupture could occur after torsion had produced the strike change. The angle of dip of the fault plane seems to be a function of brittleness. This type of faulting is unusual, and the author is not prepared to give a quantitative analysis of the mechanics of either the natural feature or his experiments. There are too many variables.

Although only the two thick beds were examined along the strike, several smaller beds were located in a 2-mile section across the strike. Since the term "bed" has been used, it would be reasonable to expect some sort of fold pattern to be traceable by connecting the beds into anticlines and synclines. There was no obvious arrangement of the known beds. Soil cover is always a problem in Piedmont geology. There may be other beds, as yet undiscovered, which are the missing limbs of anticlines and synclines.

A weak point in the argument lies in the dip changes observed in the large beds. In several places along their strikes the dips are reversed from west to east and from east to west. The explanation is probably more complex than simple folding alone. The orientation of the western bed is easily identified, since the eastern member is always the thinner. If both limbs of an anticline were exposed, this key identification would make it recognizable. The beds were not found in a reversed position at any point. This must mean that, granted that the graphite is in beds, only one limb of the fold pattern is exposed as the western bed, and the problem of the dip change remains unsolved.

The eastern bed is the widest, having an average thickness of 200 feet and a

maximum thickness of 360 feet. It is not a single, solid mass of schistose graphite but consists, rather, of graphitic schists interspersed with micaceous schist carrying no appreciable graphite. The graphitic zones average about 4 per cent graphite, although any minable width would have a much lower proportion because of the interspersed gangue material. This bed is, in general, a ridge-maker because the carbon is chemically inert and decomposes less rapidly than the more feldspathic gneisses in the region. Its low, rounded ridge is an easily marked feature, which helps make this band an exceptionally fine key horizon. The alternations of zones of graphite schist and mica schist of about equal width distinguish the eastern bed from any other.

A quarter of a mile west of the wide eastern bed there is a second characteristic key horizon. This western bed differs from the eastern in two major respects. The graphite content of the western bed varies from about 4 to a maximum of 35 per cent, whereas there is no known outcrop of the eastern bed which has more than 4 per cent graphite. In the western bed there are only two zones of graphite schist, a few feet apart. These two zones seem to be continuous. At every good outcrop of the bed, where the complete exposure is certain, both may be seen. Although neither half has a constant width, the western half is invariably the wider, averaging 40 feet in width, while the eastern half averages 19 feet. Neither varies more than a few feet from its average thickness. The mica schist separating them averages 10 feet in thickness. This tendency to maintain these uniform widths makes the western bed an unmistakable key horizon.

All the old mines and most of the prospect pits are located along the strike of the western bed. Early prospecting located all the graphite-rich zones. The old pits are still open enough to permit examination. Two old mining sites have the local names, "Old Lead Mine" and "Lead Mine Hill." The "wall-to-wall" graphite average of the former is 12 per cent; that of the latter is 35 per cent. Each of the old mining ventures failed because the ore could not be washed successfully and the quartz removed.

ORIGIN OF THE GRAPHITE

There are two possible origins for this graphite. One is magmatic and was proposed by Ebenezer Emmons[2] in 1856. He made a regional survey of the midland counties of North Carolina and noticed that the trend of the graphite was not parallel to what he considered to be the trend of metamorphics over the big area that he studied. His belief was that the graphite had come up across the foliation from below as magmatic carbon either by the reduction of CO_2 or from buried carbonates.

It has been pointed out that the strike of the foliation in the wall rock is parallel to the strike of the bands themselves as well as to the strike of the foliation of the schistose graphite. This observation is in direct opposition to that of Emmons. His conclusions must have been based on regional-reconnaissance methods rather than on strike mapping of the graphite itself. To have produced the observed structure by magmatic action, the graphite would have had to be injected along parallel planes of weakness followed by regional metamorphism. This metamorphism would have had to be so directed that the foliation developed in the country rock and graphite was parallel to the strike of the bands. Supporting evidence

[2] *Ibid.*, p. 222.

in the form of offshoots and cross-shoots, vein banding, or obvious crystalline graphite could not be found. However, it is the extreme regularity of the western bed with its two member-beds that is the major obstacle to this magmatic hypothesis. It is difficult to conceive of a set of weakness planes which would remain as separate as these two halves of the western bed over so great a distance as 10 miles. The two parts are separated by a few feet of schist, in which there is very little graphite, and that in the form of disconnected specks rather than stringers. The members themselves are distinguishable because the western one is invariably the wider. Since the case for magmatic carbon seems weak, we must consider an alternative that this graphite was the product of metamorphism acting on a carbonaceous sediment.

In seeking the mode of origin from the field evidence, it seems that the sediments must have been quite regular. The stratum which contained the carbon was probably underlain and overlain by sediments quite free from carbon. The estimate of regularity is based on the even widths of the members of the western bed. The consistent thicknesses probably indicate marine, rather than continental, sediments. The mica and quartz content of the graphite beds show that these beds were probably carbonaceous, arenaceous shales before metamorphism. Since the enclosing schists do not seem to carry any appreciable quantity of graphite, it is quite logical to suppose that these were practically free of carbon.

The beds are now found in what appears to be a folded position. The foliation of the graphite schist is parallel to the strike of the graphite beds. In order to obtain this structure it would have been necessary for the original folding which deformed the beds to continue on with the same orientation of the compressive forces until the dynamic metamorphism was complete. This is a special case and an assumption which is admittedly a weak point in the explanation of the origin. Rocks of this highly deformed nature would be expected to contain beds that marked out pitching anticlines and synclines rather than the relatively straight strike pattern which was observed. Until the entire area has been mapped and the origin of the nongraphitic gneisses and schists definitely determined, this assumption is the simplest explanation of the straight strike pattern. The dynamic metamorphic history was completed when the regular, nonplunging folds were formed and the graphite had been made from the sedimentary carbon.

Each member of each bed is believed to have been a depositional unit. Variations in the graphite content of the beds along their strike could be due to depositional irregularities. Quartz is the major mineral in all the beds. It seems to be of three different ages and to have three entirely different origins. Thin sections of these graphite schists show that some of the quartz grains are strained. These were in all probability present during the dynamic metamorphic stage and owed their deformation to this stress. The unstrained quartz occurs in two ways. Most of it is disseminated throughout the entire schist. This portion of the quartz is believed to be due to a complete *lit-par-lit* injection of much of the Raleigh area. The details and magnitude of this injection demand a more thorough investigation of the quartz of this age than can be made in the injected graphite alone. The author has never seen schist or gneiss in the Raleigh area that was not shot full of this disseminated quartz. Its injection is not comparable to the injection banding

often seen in a gneiss. There is no tendency for the quartz to be localized in bands. The rock has the appearance of having been literally stewed in quartz juice so that the quartz dispersion is homogeneous. Even those schists which carry a large percentage of graphite have a tremendous amount of quartz.

The third type of quartz occurs in stringers, small pegmatites, and veins up to 18 inches thick. This quartz is younger than any other mineralization, as the veins cut all the other structures and fill many of the joint cracks. It is believed that this third type of quartz is associated with the intrusion of granite east of Raleigh and is late Paleozoic in age.

The criteria for separating these three types of quartz are: strain for the oldest, ubiquity and lack of strain for the injected quartz, and vein structure for the youngest.

Kyanite and garnet are also found in a few places. Most of the garnet occurs in a dense veinlike zone about 6 inches wide at the Lead Mine Hill site in the western bed. The garnets preserve their dodecahedral form but are clustered closely together. The kyanite occurs in typical bladelike crystals, which definitely cut the schistosity. There is no tendency for these crystals to be aligned along the foliation. Other garnets occurring individually show that the foliation may be cut by the dodecahedron or may bend around the crystal. These garnet occurrences are believed to be contemporaneous with the metamorphism. However, the kyanite is definitely postdynamic metamorphism. The specific relationship of the kyanite age to the quartz ages has not been determined.

CONCLUSIONS

1. It is believed that these graphite beds were once carbonaceous sediments that have been folded and metamorphosed to their present condition.

2. Since graphite is chemically inert, it has not been affected by hydrothermal alteration, and the original bed structure is still recognizable. This fact will be of vital importance in deciphering the true structure of the highly metamorphosed schists and gneisses of Wake County. These rocks are so differentially altered that they cannot now be mapped as depositional units. The Raleigh graphite furnishes easily distinguishable key horizons.

ACKNOWLEDGMENT.—The author is indebted to the faculties of North Carolina State College and the University of North Carolina for assistance in interpreting these features.

DIABASE DIKES OF THE FRANKLIN FURNACE NEW JERSEY, QUADRANGLE[1]

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United States Geological Survey

ABSTRACT

Four of the numerous dikes mapped on the areal geology sheet of the *Franklin Furnace Folio*[2] cannot be included among those therein described as "mostly basic, including nepheline tinguaite, leucite tinguaite, and camptonite" of "post-Ordovician" age. Field and petrographic study and chemical analyses show that these four dikes form a closely related group, sharply distinguishable from all other known dike rocks of the area. They are in all essential respects similar to the Triassic diabases of other parts of New Jersey and the Atlantic seaboard states, and there is no reason to question their Triassic age.

LOCATION OF DIKES

The locations of the four dikes are as follows:

1. East of Howell Pond, near the locality marked "Pinckneyville," in the southwestern part of the quadrangle. This occurrence will be called the "Pinckneyville dike" in the following discussion. On the rectangular co-ordinate grid system of the Sussex County, N.J., *Base Map* (N.J. Highway Dept.–Federal Works Agency, Public Roads Administration, 1939), the dike extends for about ¼ mile N. 45° E., approximately diagonally across the square whose co-ordinates are 1991–92 and 804–5. The length of the dike is at least twice that indicated on the *Folio* map. The dike forms the topographic ridge and probably extends somewhat farther to the east than is indicated on that map.

2. About 2 miles east-northeast of the Pinckyneyville dike, in the vicinity of Sussex Mills. Here this is called the "Sussex Mills dike." It extends in a northeasterly direction for about ⅛ mile, as indicated on the *Folio* map, and is about 30 feet wide. Its co-ordinates are 2000–2001, 805–6.

3. Near the middle of the southern edge of the quadrangle, about a mile south of the old Mount Sherman prospect. This is called the "Mount Sherman dike," and its co-ordinates are 2011–12, 793–94. It is the smallest of the four dikes, being only some 10 feet long and a few feet wide. Its strike, approximately N. 50° W., is different from that of the others as shown on the *Folio* map. It is about 200 feet east of a farmhouse.

4. Near the hamlet of Russia, in the southeast corner of the quadrangle. Its co-ordinates are approximately 2035–36 and 804–5. It extends for some 200 feet, about N. 80° E., along both sides of the road. It has been freshly blasted along the roadside and offers good exposures.

OTHER DIKES

All the other known post-Ordovician dikes of the quadrangle intrude either pre-Cambrian marble (Franklin limestone) or Cambrian and Ordovician rocks, whereas the diabase dikes cut pre-Cambrian gneiss—the Pinckneyville dike being in the Pochuck gabbro gneiss, while the Sussex Mills, Mount Sherman, and Russia dikes are in the Losee diorite gneiss. All four diabase dikes occur in the southern third of the quadrangle, remote from the other known dike rocks of the northern half of the quadrangle. Whether

[1] Published by permission of the director, United States Geological Survey.

[2] *U.S. Geol. Surv. Geol. Atlas, Franklin Furnace Folio 161* (1908).

similar diabasic intrusives occur to the south, in the Lake Hopatcong quadrangle, is not known; but at least through 1944 no Triassic rocks, igneous or sedimentary, had been recorded in either Sussex or Warren County.[3] Near the extreme southern border of Morris County, however, where Morris, Somerset, and Hunterdon counties adjoin, an isolated diabase dike is indicated on the geological map of New Jersey (J. Volney Lewis and Henry B. Kümmel, 1910–12). It is about 6 miles north of the belt of Triassic sedimentary and igneous rocks crossing the state and fully 15 miles from the Franklin Furnace quadrangle. It is mapped as intruding Byram gneiss and trends northeast-southwest (conforming with the regional structure). It is about $\frac{1}{3}$ mile in length.

A careful examination has been made of all mapped dikes in the Franklin Furnace quadrangle, as well as of a great number that have not been mapped, and none of them, except the four herein described, is diabase. No Triassic sedimentary rocks are known in the region, and the wide Triassic belt of sedimentary and igneous rocks crossing New Jersey is 20 or more miles to the southeast. Nevertheless, the occurrence of these four dikes is believed to be of significance in that it indicates that the locus of the alkalic Beemerville nepheline syenite and associated volcanics is at no great distance from the outliers of the geotectonic region of Triassic faulting and vulcanism. In a paper now in preparation, other instances of association of nepheline syenite and known Triassic igneous rocks will be discussed.

In a third paper four other bodies of intrusive rock forming a closely related group, markedly different from both the Beemerville alkalic suite and the presumably Triassic diabases herein described, and designated the "Franklin Furnace Kugel-minettes," will be discussed. On the sketch map (Fig. 1) are shown the geographical relations of the diabase dikes, the Beemerville alkalic complex, the Kugel-minettes, and the northern boundary of the New Jersey Triassic sedimentary-igneous belt, which is indicated by the heavy dashed line.

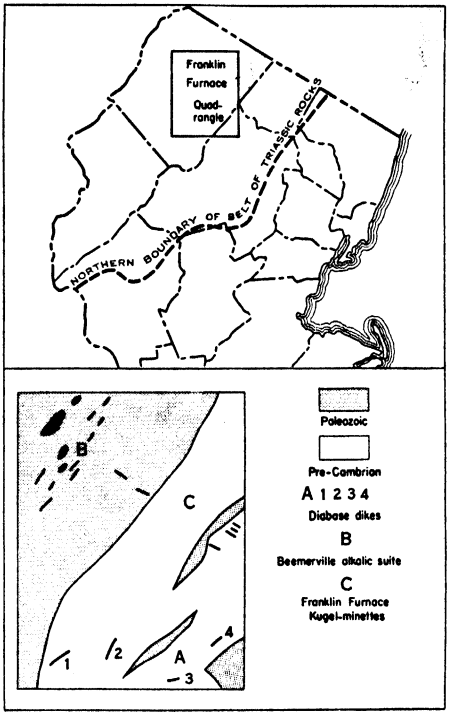

FIG. 1.—Map showing diabase dikes in the Franklin Furnace quadrangle.

FIELD RELATIONS OF THE DIKES

The Pinckneyville dike is most easily reached from the road east of Howell's Pond. It forms massive ledges, and its contacts with the intruded Pochuck gabbro gneiss show no observable contact-metamorphism.

[3] Personal communication from Meredith E. Johnson, state geologist of New Jersey, December 14, 1944.

The Sussex Mills dike is similar in topographic expression to the Pinckneyville dike; it, too, forms a low ridge which, with a few gaps, can be followed for many hundred feet; in a few places contacts with the gneiss can be seen. The diabase varies irregularly from fine grained to rather coarse grained. Probably the difference is due to a chilling effect.

The Mount Sherman dike is without any topographic expression and is noticeable only by its contrast in appearance with the light-colored gneiss, across whose foliation it cuts sharply. It has a dull, stony aspect.

The Russia dike differs from the others in several respects. It has undergone some folding, and superficially resembles a greenstone and, without detailed study, might be thought part of the pre-Cambrian metamorphic complex. Contorted veinlets with much epidote cut the rock, and the dike is distinctly schistose, and slickensiding can be seen. No contacts with the local light-colored gneiss were found.

PETROGRAPHY OF THE DIABASES

The diabasic rocks of the Atlantic seaboard are generally known to be fairly uniform in mineralogical and chemical composition, and these four occurrences are no exceptions to the rule.

The Pinckneyville dike is a fine-grained, dark-gray rock, of rather uniform texture throughout its exposure, showing little or no weathering. Grains of pyrite, 1–2 mm. across, are perfectly fresh and show no signs of oxidation.

The effect of rapid cooling, on contact of the dike with the enclosing granite gneiss, is evident on microscopic study. The slowly cooled rock is holocrystalline, with excellent development of all the major minerals, whereas, in the chilled border phases, an obscurely crystallized groundmass contains isolated well-developed early-formed crystals. Intermediate plagioclase, diopsidic pyroxene, quartz, and black ore grains comprise the bulk of the well-crystallized diabase. The quartz usually occurs in isolated rounded areas, often with pale-green chlorite. The ore grains commonly show a characteristic skeletal development. In very minor quantity are found sphene, always of opaque grains, and apatite. In contrast, the chilled border phase shows no quartz or accessory minerals; only a crowded semi-glassy brown groundmass in which euhedral plagioclase crystals and diopsidic pyroxene, in single crystals and complexly intergrown composite groups, are strewn.

A thin section across the contact of the diabase and the granite gneiss shows a hair-line contact, with no suggestion of any contamination of the diabase by the adjoining gneiss. The gneiss, which consists essentially of sericitized microcline and quartz, likewise shows no indication of mechanical deformation by the intrusive diabase.

The Sussex Mills dike consists of two main varieties, which, however, grade into each other, one of distinctly fine grain, the other of coarser. There is also a variety in which hydrothermal alteration processes have caused marked changes, producing a rock rich in quartz and epidote, similar to the Russia rock described below.

The fine-grained Sussex Mills rock is dense, with tiny feldspar cleavage surfaces and minute pyrite grains breaking up the dull appearance of the fresh rock surface. Weathering is almost absent. The other variety is distinctly coarser, with a granularity suggesting a gabbro. Pyrite and cleavage surfaces of feldspar and of pyroxene grains are visible. The microscope shows, however, that the

mineralogy of the two rocks is very similar. The analysis given below is of the fine-grained rock.

The finer-grained variety of the Sussex Mills rock shows, under the microscope, prophyritic crystals of perfectly fresh calcic plagioclase and pyroxene, in an ill-defined yet holocrystalline groundmass consisting of these two minerals, with subordinate amounts of ilmenite-magnetite, leucoxene, chlorite, biotite, and pyrite. The groundmass is characterized by a rectangular network of ilmenite-magnetite crystallites, which are not present in the coarser variety.

The coarser-grained variety of the Sussex Mills rock consists mainly of pyroxene and calcic plagioclase, with subordinate quartz and sodic plagioclase (micropegamatite), also apatite, ilmenite-magnetite, leucoxene-sphene, and indefinite chloritic areas. With these last are some wisps of green hornblende or brown biotite. Microscopic fractures, filled with green chlorite, traverse the section. The calcic plagioclases show a slight turbidity due to sericitization.

The difference between the two varieties is presumably due to different conditions of crystallization; the fine-grained rock being chilled after the development of the porphyritic plagioclases and pyroxenes and the magnetite-ilmenite grid structure. In the coarser-grained variety there was a relatively slow cooling, sufficient to permit a uniform crystallization, and agglomeration of the grid-material into equant grains.

The hydrothermally altered variety of the Sussex Mills dike is a dull greenish-gray stony rock, seamed with green bands of epidote-quartz, which project from the weathered surface. Slickensides are preserved on these harder bands. Microscopically, the typical diabase texture has been obliterated; no pyroxenes remain and but little of the calcic plagioclase. Most of the feldspar present appears to be clear, not turbid, is untwinned, and near oligoclase-andesine in composition, judging by the refraction. Clear quartz, with much yellowish epidote, is abundant. Some of the epidote enclosed in quartz is euhedral. Green chlorite is abundant in the quartz-epidote areas. Leucoxenized titanited grains are strewn about the section, and there is some calcite.

This rock shows clearly the relation between the very similar Russia rock and the other diabases. Without it there might be some doubt as to whether the Russia rock should be considered a diabase. Finding this altered diabase also raises the question as to whether certain other rocks of diabasic character, referred on the basis of alteration to an age older than Triassic, may not actually be Triassic.[4]

The Mount Sherman dike, though only 2 feet wide, nevertheless shows a contrast between the middle coarser part and the finer-grained chilled borders. An interesting feature is the presence of numerous masses of bright-green serpentine or chlorite, whose sharp outlines strongly suggest replacement of olivine. The reddish proxene is, in general, completely unaltered, but a few grains contain a core of the green mineral. These replacements are present in both the coarse- and the fine-grained rock. In the latter there is a distinct porphyritic development, mostly of plagioclase, with less pyroxene; the groundmass is an obscure aggregate of extremely fine grain, probably, in part, glass. Microscopic veinlets that appear to be of the same substance that has replaced the (possible) olivine also traverse the rock. No quartz was seen.

[4] Anna I. Jonas, "Pre-Cambrian and Triassic Diabase in Eastern Pennsylvania," *Bull. Amer. Mus. Nat. Hist.*, Vol. XXXVII (1917), pp. 173–81.

The Russia dike is uniform, of fine grain, gray-green in color, and is fresh and tough. Thin veinlets of quartz-epidote cut the dike, which is somewhat schistose and slickensided. The contact with the enclosing light-colored gneiss was not observed. Except for a few specks of pyrite, no minerals can be distinguished in the hand specimen. Microscopically, it resembles the coarse variety of the Sussex Mills dike, except that little or no quartz is present. However, the Russia dike shows more alteration effects than any of the others. There are no porphyritic plagioclase crystals; the pyroxene, which is distinctly red, shows peripheral replacement by wispy amphibole; and much of the coarse ilmenite has gone over to leucoxene.

TABLE 1*

CHEMICAL ANALYSES OF THE DIABASE DIKES FROM THE FRANKLIN FURNACE QUADRANGLE

(Charles Milton, Analyst)

	A	B	C	D
SiO_2	49.25	49.49	49.03	47.42
Al_2O_3	13.62	14.63	13.72	13.99
Fe_2O_3	4.22	5.39	4.11	1.97
FeO	9.18	8.83	10.44	11.49
MgO	4.43	4.62	5.57	6.62
CaO	7.63	8.03	9.96	7.81
Na_2O	2.61	2.70	1.90	3.30
K_2O	2.00	1.60	0.81	1.54
H_2O^+	2.16	1.56	1.89	2.71
H_2O^-	0.07	0.10	0.06	0.11
CO_2	.....	.....	.....	trace
TiO_2	3.31	2.56	2.43	2.30
P_2O_5	1.23	0.46	0.08	0.30
S	0.18	0.23	0.09	0.26
MnO	0.26	0.20	0.23	0.15
BaO	0.02	0.03	0.02	0.00
ZrO_2	.....	.....	.....	0.05
Total	100.17	100.43	100.34	100.02
Density	2.92	3.01	3.00	3.06

* A, diabase, Pinckneyville, Andover Township, Sussex County, New Jersey; B, diabase, Sussex Mills, Sparta Township, Sussex County, New Jersey; C, diabase, Mount Sherman, Sparta Township, Sussex County, New Jersey; and D, diabase, Russia, Jefferson Township, Morris County, New Jersey.

These four analyses are very similar and bear a close resemblance to published analyses of Triassic diabases from other localities. In the C.I.P.W. classification, the norm of the Pinckneyville, Sussex Mills, and Russia diabase dikes is III.5.3.4; the norm of the Mount Sherman differs but slightly, being III.5.4.4; this difference reflects a slightly higher ratio of CaO to alkalies.

The norms of the Pinckneyville and Sussex Mills rocks are very similar, as the petrographic description of these would indicate. The Mount Sherman rock shows the highest An–Ab ratio, presumably correlated with the small size of the dike. Rapid chilling would produce a fine-grained compact rock, which might be less susceptible to albitization or similar deuteric effects. The Russia rock, on the other hand, has the lowest An–Ab ratio, and it is this dike (with the altered phase of the Sussex Mills dike) which shows the most effect of hydrothermal activity, such as destruction of the early-formed calcic-plagioclase crystals, formation of quartz-epidote veinlets, and formation of leuxocene and pyrite.

TABLE 2

NORMS OF THE DIABASE DIKES FROM THE FRANKLIN FURNACE QUADRANGLE

	Norms			
	Pinckneyville (III.5.3.4)	Sussex Mills (III.5.3.4)	Mount Sherman (III.5.4.4)	Russia (III.5.3.4)
Quartz	5.6	4.1	4.9	.....
Orthoclase	11.7	9.5	5.0	8.9
Albite	22.0	23.1	15.7	27.8
Anorthite	19.5	22.8	26.7	18.9
Diopside	8.2	11.7	18.7	15.8
Hypersthene	14.8	13.2	16.3	.....
Olivine	.....	.....	.....	17.3
Magnetite	6.0	7.9	6.0	2.8
Ilmenite	6.1	4.9	4.6	4.4
Water, etc.	5.2	3.0	2.3	3.9
Total	99.1	100.2	100.2	99.8

ACKNOWLEDGMENTS.—W. T. Schaller, Earl Ingerson, and C. S. Ross, of the Geological Survey, have kindly read and instructively commented on this manuscript.

REVIEWS

Dating the Past: An Introduction to Geochronology. By FREDERICK E. ZEUNER. London: Methuen & Co., Ltd., 1946. Pp. xviii+444; figs. 101; pls. 24. 30*s*.

Dating the Past is a companion volume to Zeuner's *The Pleistocene Period*. While the latter deals mainly with the evidence for the relative and absolute chronology of the Pleistocene, the present book is concerned with the various methods of dating the past, with geochronology in general, and with its application to archeology in particular (p. v). Geochronology is defined as "the science of dating in terms of years those periods of the past to which the human historical calendar does not apply" (p. 5). The word was probably first employed in this sense by Gerard De Geer in 1912 with reference to his measuring time by means of the late-glacial and postglacial varved clays.[1] Knowledge of age and time is necessary for perspective and understanding, as is reflected in the layman's ever recurring questions: How old? How long ago? How long a time did it take?

Zeuner treats four geochronological means (p. 5) and divides the treatise into four corresponding parts, viz.: (1) tree-rings: dating early history and late prehistory, especially in North America (pp. 6–19); (2) varved clays: dating the Metal ages, New and Middle Stone ages, and the climatic phases which followed the Ice Age (pp. 20–109); (3) the varying distribution of the solar radiation: dating the Old Stone Age, the phases of the Ice Age and the Pluvial phases of the warmer countries (pp. 110–304); and (4) radioactive disintegration: dating the history of the earth and life before the arrival of man (pp. 305–85).

The expressions "tree-ring analysis," which has been employed by others, and "varved clay analysis" and "varve analysis," which are probably new (pp. 5, 6, 20), seem improper since for use in dating and climatic research tree-rings and varves are not resolved into their constituent parts. "Tree-ring research," "tree-ring science," "varve research," and "varve science," or "varvology," seem more appropriate.

Tree-rings have long been studied from many angles. Their chronological importance is based on the fact that their width is notably influenced near the dry limit of tree growth by the water supply available during the growing season and near the cold limit by the temperature of the season of growth. The modern ring research has two distinct aims: (1) dating of prehistoric ruins (by means of wooden beams), and (2) investigation of past variations in climate, especially in rainfall. The study as a whole is usually called "dendrochronology," and "tree-ring hydrology" and "dendroclimatology" are taken as subdivisions. The first method of using tree-rings in archeological dating and the most-used method in climatic studies was devised by A. E. Douglass of the University of Arizona. This method is outlined with reference to Waldo Glock's detailed description. Some examples of ruin-dating in the American Southwest are given, and cycles in radial growth, especially the eleven-year, sun-spot cycle, are discussed (pp. 12–19). It should be added that Harold S. Gladwin has developed another method of tree-ring correlation for beam-dating[2] and also that Edmund Schulman of the University of Arizona has carried out important dendroclimatic research.

Varve is the distinguishable annual deposit of any sediment. It is the equivalent to the annual ring in wood. Varves occur in both organic and inorganic sediments. True or supposed varves in various pre-Pleistocene sediments are listed on pages 36–38. However, only three kinds are of chronological importance so far, viz., the glacial varved clay, the postglacial varved silt of northern Sweden, and the varved gyttja in Faulensee, Switzerland, too recently described to be considered by Zeuner.[3]

[1] "A Geochronology of the Last 12,000 Years," *Compt. rend. XIe Cong. géol. internat.* (Stockholm, 1912), pp. 241–53.

[2] "Tree-Ring Analysis: Methods of Correlation," *Medallion Papers No. XXVIII* (1940), Gila Pueblo, Globe, Ariz. Reviewed in *Amer. Anthropologist*, Vol. XLVIII (1946), pp. 433–36.

[3] Max Welten, "Pollenanalytische, stratigraphische und geochronologische Untersuchungen aus

The study of varves was originated by De Geer in 1878, and practically all phases of the method in use were developed by him.

The postglacial varved silt is still being formed, but since its latest beds are below the level of the Baltic, the count, which comprises 7,760 varves, does not extend up to the present. However, by means of a careful chronological study of the land rise, Ragnar Lidén found that the last measured varve was deposited about A.D. 920; and the first postglacial varve was thus laid down about the year 6840 B.C. (p. 26).

The postglacial varved clay is conformably underlain by glacial varved clay, which has enabled a direct backward extension of the varve chronology. This glacial chronology embraces the wide belt of ice release from northeastern Scania (56° N.), the southernmost province of Sweden, to Stugun in northern Sweden (63°10′ N., 15½° E.). In the opinion of Sigurd Hansen and of the reviewer it cannot be extended over southwestern Scania and the Danish Islands (cf. pp. 27, 29, 31). Not all data have been published, but this glacial-varve chronology probably comprises 4,500 to 5,000 years. Consequently, the last ice border stood in northeastern Scania some 13,500 years ago; and 13,500 years is the length of the combined postglacial and late-glacial Swedish varve chronology.

In Sweden the bisection of the ice remnant at Stugun and the accompanying change from glacial to postglacial varved clay is, after De Geer's suggestion, taken as the demarcation line between the Glacial and the Postglacial. So defined, the Postglacial began in 6840 B.C. and comprises the last 8,800 years (p. 28). Other suggested demarcation lines—the beginning of ice retreat from the Fennoscandian moraines (p. 28), the attainment of the modern temperature level in the southern and greater parts of the glaciated regions,[4] and the arrival of the modern forest in Denmark[5]—are based on recognizable climatic changes for long-distance correlation and broader application. They place the transition at 9000–10,000 B.P. (before the present).

Matti Sauramo's varve studies in Finland and mine in North America are summarized (pp. 32–35). Since two figures and statements attributed to me are incorrect, my chronology[6] of the ice retreat from Mattawa, Ontario, to James Bay is here restated: 1,000 years estimated + 2,000 counted of retreat + 2,000 in part counted of oscillations in the Cochrane region + 1,000 years estimated for final retreat to James Bay: total 6,000 years. The subsequent postglacial age is, as just stated, taken to be 9,000 years, and the ice border at Mattawa is so given the age of 15,000 years. The important Mattawa ice border can also be dated thus: The Cochrane ice-border oscillations and moraines are probably correlatives of the Fennoscandian moraines, now dated by Sauramo at about 8900–8000 B.C., or roughly 11,000–10,000 B.P. If the drainage of Lake Ojibway-Barlow, which just preceded the first ice readvance to and beyond Cochrane, therefore be dated at 11,000 B.P., and if 2,000 counted and 1,000 estimated years be added, the date of the border at Mattawa will be 14,000 B.P.

The application of the varved-clay chronology to archeological dating is usually by a round-about route (p. 46). Some stages and shorelines of the Baltic have been directly geochronologically dated, say within one hundred years. And by means of pollen grains in varve series the regular appearance, or the maximum frequency, of characteristic forest trees as well as some distinct climatic ages have been rather exactly dated. For instance, the influx across central Sweden of salt water into the Baltic basin which introduced the Yoldia stage took place from 8000 to 7900 B.C., the transition to the Ancylus stage about 6500 B.C., and that to the Littorina stage about 5000 B.C. (pp. 51–53). The alder became a regular constituent of the forest about 6350 B.C., and the spruce between 1000 and 950 B.C.; and the birch reached it greatest frequency about 4200 B.C. The temperature was (on 63°10′ N.) highest during the age 5000–3500 B.C., and it dropped distinctly around 1000 B.C.[7]

With the help of such starting-points other Baltic shorelines and stages and other pollen

dem Faulenseemoos bei Spiez," *Veröff. Geobotan. Inst. Rübel in Zürich*, No. 21 (1944). Reviewed in *Amer. Jour. Sci.*, Vol. CCXLIV (June, 1946), pp. 442–47.

[4] Ernst Antevs, "Late-glacial Correlations and Ice Recession in Manitoba," *Geol. Surv. Canada, Mem. 168* (1931), pp. 2, 6.

[5] Johs Iversen, "Plantevaekst, Dyreliv og Klima i det senglaciale Danmark," *Geol. Fören. i Stockholm Förhandl.*, Vol. LXIX (1947), p. 68.

[6] Pp. 18, 19, 29, 33, 34 of ftn. 4.

[7] Erik Fromm, "Geochronologisch datierte Pollendiagramme und Diatoméenanalysen aus Ångermanland," *Geol. Fören. i Stockholm Förhandl.*, Vol. LX (1938), pp. 365–81; see pp. 379–80.

horizons and pollen profiles have been dated approximately and so have archeological finds associated with shores or located in peat bogs. In the last instance the possibility of dating is usually dependent on climatic changes enabling correlation.

The history of the Baltic during the last 11,000 years is outlined mainly after Sauramo's summaries (pp. 47–55). The pollen analysis and other botanical ways of studying the changes of the forest composition and thus of the climate are described as well as the deduced climatic stages (pp. 56–72). A long chapter reviews important archeological sites from the past 15,000 years in northwestern Europe (pp. 72–109). The use of the floral and climatic succession as basis for chronology beyond Fennoscandia meets two well-known complications, viz., the distinct zoning and the northward migration of the zones as the ice border withdrew and the varying rate of spread of the species (p. 104). However, dated, distinct, and widespread climatic changes and geological events enable at least many approximate correlations and datings (p. 108).

Part III, "Dating the Old Stone Age, the Phases of the Ice Age and the Pluvial Phases of the Warmer Countries," begins with a chapter on the relative and absolute chronology of the Pleistocene (pp. 110–45), which is a summary of Zeuner's book *The Pleistocene Period*, reviewed in *Journal of Geology*, Volume LV (1947), pages 446–50. The absolute astronomic chronology of the glacials and interglacials shows a correspondence with the relative, stratigraphic-climatic chronology which is difficult to explain as mere chance. However, "if anybody regards the application of the absolute time-scale as unconvincing, it is open to him to discard this part of the Pleistocene chronology altogether. That, however, will not effect the value of the relative, climatic, chronology" (p. 110). While the astronomic chronology might really apply, still, as was pointed out in the above-mentioned review, not every pronounced minimum of summer insolation, especially not the first in the group of two or three, need have been represented by a continental glaciation; and, as a consequence, some of Zeuner's dates perhaps need adjustments. For instance, ice sheets were probably not formed upon the radiation minimum 115,000 B.P., that is, there was probably no Würm, or LG1, continental glaciation, but the Warthe (and the Iowan) may, rather, represent the radiation minimum of 72,000 B.P.

The dating of the Paleolithic finds and sites in temperate Europe is made on geologic and/or paleontologic grounds and thus fitted into the astronomic chronology. It is not made with implements used as characteristic fossils (p. 146). Industries dated in this manner have been used in building up relative and absolute chronologies of the Paleolithic of central Europe, France, and Britain (p. 200). If, as suggested, the first ice sheets of the Last Glacial were developed upon the radiation minimum of 72,000 B.P., the Mousterian culture of the Neanderthal man which belongs to the Last Interglacial and the next stage of extensive glaciation would be considerably extended upward in the time scale to, say, 50,000 or 40,000 B.P., which would be in accord with archeology.[8]

Other chapters deal with the "Chronology of the Pleistocene and Palaeolithic of the Mediterranean Area" (pp. 201–43), "Climatic Phases, Early Man and Human Industries in Africa, Asia, Australia, and America" (pp. 423–81), and "Chronology of Early Man and His Cultures" (pp. 281–304). The *Pithecanthropus* group of *Homo* is believed to have lived prior to 400,000 B.P. (p. 295), the Neanderthal group during the Last Interglacial and the first continental glaciation of the Last Glacial (p. 296), and the *Homo sapiens* group may have existed at least during the last quarter-million years (p. 298). The Neanderthal man produced the Mousterian flake industry and *Homo sapiens* the blade cultures and perhaps the hand-ax industry (p. 302).

Part IV, "Dating the History of the Earth and of Life before the Arrival of Man," begins with a chapter on the measurement of geological time previous to the Pleistocene Ice Age (pp. 305–33). It reviews the methods of estimating and measuring time based on organic evolution, sedimentation, accumulation of salts in the oceans, cooling of the earth, and radioactive decomposition, which last method has superseded the others. The oldest radioactive minerals are about 1,750,000,000 years old, and Arthur Holmes has estimated the earth's minimum age at 1,900,000,000–2,000,000,000 years. Of this about three-fourths belong to the Pre-Cambrian, the Cambrian beginning some 500,000,000 years ago (pp. 333, 335, 337). The dates serve as bases for estimates of the time rates of

[8] Robert J. Braidwood, "Geochronology, and Geological Factors Relative to Man's Past," pp. 43–56 in *Human Origins: Selected Readings* (2d ser.; Chicago: University of Chicago, 1946). See p. 52.

weathering, erosion, mountain-building, and other geological processes (pp. 337–49).

The last chapter, "Biological Evolution and Time," stresses the significance of geochronology for studies on the evolution of life (pp. 350–85). In a table the minimum time of existence of several animal groups is given, which for man is 1,000,000 years, for manlike apes 50,000,000, and for mammals 160,000,000 years (p. 351). The organic evolution has proceeded at a varying pace. It has been fast and has produced many new types within relatively short spaces of time (or perhaps 50,000,000 years), then slow during longer ages (pp. 352–62). While subspecies have developed within the last 7,500 years (pp. 364–66), full species of terrestrial animals ranging from mammals to insects have required at least 500,000 years and species of aquatic animals still longer time (pp. 375, 378). "With the establishment of geochronological time-scales a new element has entered into the study of evolution and . . . . , as these time-scales are improved, they are bound to become increasingly valuable as measures of the actual rates of evolution" (p. 384).

This is a very impressive book, and the seven years spent in its preparation have clearly been years of intensive research and deliberation. It should serve as a useful handbook and as an inspiration to students of several sciences dealing with the past.

ERNST ANTEVS

Glacial Geology and the Pleistocene Epoch. By RICHARD FOSTER FLINT. New York: John Wiley & Sons, Inc.; London: Chapman & Hall, Ltd., 1947. Pp. 589; pls. 6; figs. 88. $6.00.

The appearance of this book by Professor Flint of Yale University marks a milestone in the history of Pleistocene studies in that it is the first formal text by an American author to cover the fields of glacial and Pleistocene geology. In addition, it provides a comprehensive summary of a vast literature and a means of evaluating problems on which further research is needed. So many fundamental problems are discussed that a full review will not be attempted.

About one-third of the text concerns glacial processes, land forms, and deposits, while most of the remainder is devoted to Pleistocene history. There are six folded plates which show: existing glaciers, storm tracks, pack ice, and selected ocean currents in (1) the northern hemisphere and (2) the southern hemisphere; (3) inferred glaciers, pack ice, and storm tracks in the northern hemisphere during the maximum of Wisconsin glaciation; (4) distribution of loess and Urströmtäler in Europe; (5) glaciated areas in western United States; and (6) glaciated areas of Europe. Numerous additional glacial maps occur as figures in the text. The text figures include seventy maps and diagrams but only eighteen photographs. About nine hundred references, a high percentage of them cited in the text, are listed separately. There is full coverage of important foreign titles.

In an introductory chapter the historical background is briefly summarized. This is followed by nine chapters on glacial geology dealing with the following topics: glacier ice and glacier motion, regimen of glaciers, existing glaciers, glacial erosion, glacial deposits and their interpretation, glacial drainage, ice-thrust features, and eolian deposits. The chapter on existing glaciers covers the entire world and is analytical rather than descriptive. In the treatment of glacial deposition the close relations between till and stratified deposits are emphasized. T. C. Chamberlin's terms, "dumping," "pushing," and "lodgement," are revived for the three principal processes in till deposition, and distinction is made between "basal till" formed by lodgement and "superglacial till formed by ablation.

The subject of Pleistocene history is introduced by a chapter on glacial stratigraphy in which full consideration is given to criteria for the differentiation of drift sheets, the problem of the Pliocene-Pleistocene boundary, and Pleistocene nomenclature. It is suggested that the terms "Tertiary" and "Quaternary" be dropped and that "recent" and "postglacial" be used only in an informal sense. The Pleistocene is considered an epoch and is provisionally defined on the basis of glacial climates and fossil vertebrates to include all of post-Pliocene time. The classification for the North American Pleistocene is given in Table 1.

In the next six chapters there follows a concise review of the Pleistocene record throughout the world. This section is more than an up-to-date regional summary in that it includes significant original interpretations by the writer. Comparisons between foreign glacial features and similar features in North America are especially helpful. The final chapter in the section considers methods of estimating Pleistocene time.

The concluding chapters in the book deal with: changes in level of land and sea, glacial

and interglacial climates, postglacial climates, causes of glacial climates, and the Pleistocene fossil record. The causes of glacial climates are narrowed to three groups of hypotheses based on: (1) topographic changes, (2) periodic planetary changes, and (3) variations in solar heat. A "solar-topographic hypothesis," offered by the writer, is a modification and extension of older topographic hypotheses and Simpson's solar hypothesis. The Milankovitch hypothesis, advocated by F. E. Zeuner in two recent books on the Pleistocene (*The Pleistocene Period: Its Climate, Chronology and Faunal Successions*

TABLE 1

Era	Period (System)	Epoch (Series)	Age (Stage)	Sub-age (Substage)
Cenozoic	Cenozoic	Pleistocene	Wisconsin	Mankato Cary Tazewell Iowan
			Sangamon Illinoian *Yarmouth* Kansan *Aftonian* Nebraskan	
		Pliocene		

[1945], and *Dating the Past: An Introduction to Geochronology* [1946], is opposed on seven different counts.

In looking back over the contents of the book one is impressed by the years of compilation and careful consideration along with constant revisions that must have preceded the final copy. All students of the Pleistocene, and others in allied fields, welcome it as an invaluable text and reference book.

LELAND HORBERG

Geological Expedition to the Lesser Sunda Islands under the Leadership of H. A. Brouwer. Vols. II, III, and IV. Amsterdam: Noord-Hollandische Uitgevers Maatschappij, 1940–42. 17.00, 17.00, and 22.50 gulden.

The first volume of this four-volume series was well reviewed in considerable detail by Carey Croneis in 1940 (*Journal of Geology*, Vol. XLVIII, pp. 553–56). Attention should now be called to the three succeeding volumes whose appearance during the war has completed the set of formal reports prepared by the members of this important expedition under the leadership of Dr. Brouwer. They extend and round out the studies whose partial presentation in Volume I in 1940 aroused so much interest. Dr. Brouwer and his colleagues should receive the belated appreciation of all geologists interested in this portion of the globe which reveals so much that bears vitally on various diverse geologic problems. Timor, in particular, because of its exceptional features, has aroused hopes in not a few geologists that they may sometime have an opportunity to visit it.

Volume II, of 395 pages, presents the following contributions: H. A. Brouwer, "Geological and Petrological Investigations on Alkali and Calc-Alkali Rocks of the Islands Adonara, Lomblen and Batoe Tara"; W. P. de Roever, "Geological Investigations in the Southwestern Moetis Region (Netherlands Timor)"; J. H. van Voorthuysen, "Geologische Untersuchungen im Distrikt Amfoan (Nordwest Timor)"; and C. Wanner, "Neue Permische Lamellibranchiaten von Timor."

Volume III, of 380 pages, includes three reports: F. P. van West, "Geological Investigations in the Miomaffo Region (Netherlands Timor)"; D. L. de Bruyne, "Sur la composition et la genèse du bassin central de Timor"; and J. D. de Jong, "Geological Investigations in West Wetar, Lirang and Solor."

Volume IV, 423 pages, is in six sections: J. Heering, "Geological Investigations in East Wetar, Alor and Poera Besar"; C. Wanner, "Neue Beiträge zur Gastropodenfauna des Perm von Timor"; W. P. de Roever, Olivine-Basalts and Their Alkaline Differentiates in the Permian of Timor"; H. A. Brouwer, "Granodioritic Intrusions and Their Metamorphic Aureoles in the Young-Tertiary of Central Flores"; J. D. de Jong, "Hydrothermal Metamorphism in the Lowo Ria Region, Central Flores"; and H. A. Brouwer, "Summary of the Geological Results of the Expedition."

The concluding summary report of 50 pages by Brouwer will be found not only helpful as a general account of the geology of these islands but also particularly valuable for its able treatment of the larger tectonic and magmatic problems and their interrelationships. Few areas of comparable size can contribute more to our understanding of the earth's diastrophic behavior.

R. T. C.

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September 25, 1947

Dr. R. T. Chamberlin, Editor
Journal of Geology
5750 Ellis Ave.
Chicago 37, Illinois

DEAR DR. CHAMBERLIN:

Finland has an excellent and keenly scientific-minded Technical Institute, Teknillinen Korkeakoulu. During the war its library was bombed and totally destroyed.

On my recent trip to Finland for the American Friends Service Committee, I discussed the situation with Dr. Martti Levon, Director of the Institute. He said he would welcome gifts of scientific and technical books and periodicals from America to take the place of those destroyed. In the remarkable efforts for recovery that the Finns are making, the lack of technical library facilities is a very serious handicap. It would be a practical act of friendship to a nation that holds America in high regard if Americans should contribute good technical books and periodicals to this library.

Any such gifts should be marked for the Institute of Technology, Helsinki, and sent to the Legation of Finland, 2144 Wyoming Ave., N.W., Washington, D.C. Dr. K. T. Jutila, the Finnish Minister, will arrange for their being shipped to Finland.

Sincerely yours,

ARTHUR E. MORGAN

AMERICAN FRIENDS SERVICE COMMITTEE
YELLOW SPRINGS, OHIO

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